JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 99, NO. B3, PAGES 4293-4221, MARCH 10, 1994

Formation and emplacement of the Josephine ophiolite and the Nevadan orogeny in the Klamath Mountains, California-Oregon: U/Pb zircon and $^{40}Ar/^{39}Ar$ geochronology

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Cordilleran ophiolites typically occur as basement for accreted terranes. In the Klamath Mountains, ophiolitic terranes were progressively accreted by underthrusting beneath North America. The Josephine ophiolite is the youngest of the Klamath ophiolites and forms the basement for a thick Late Jurassic flysch sequence (Galice Formation). This ophiolite-flysch terrane forms an east dipping thrust sheet sandwiched between older rocks of the Klamath Mountains above and a coeval plutonic-volcanic arc complex below. The outcrop pattern of the roof (Orleans) thrust indicates a minimum displacement of 40 km, and geophysical studies suggest >110 km of displacement. The basal (Madstone Cabin) thrust is associated with an amphibolitic sole and has a minimum displacement of 12 km. A rapid sequence of events, from ophiolite generation to thrust emplacement, has been determined using $^{40}Ar/^{39}Ar$ and Pb/U geochronology. Ophiolite generation occurred at 162-164 Ma, a thin hemipelagic sequence was deposited from 162 to 157 Ma, and flysch deposition took place between 157 and 150 Ma. Tight age constraints on thrusting and low-grade metamorphism associated with ophiolite emplacement (Nevadan orogeny) are provided by abundant calc-alkaline dikes and plutons ranging in age from 151 to 139 Ma. Deformation and metamorphism related to the Nevadan orogeny appears to have extended from ~155 to 135 Ma. Most of the crustal shortening took place by thrusting, constrained to have occurred from ~155 to 150 Ma on both the roof and basal thrusts. Minimum rates of displacement are 2.4 and 3.6 mm/year for the basal and roof thrusts, respectively, but correlations with coeval thrusts yield rates of 8.4 and 22 mm/year (within the range of plate velocities). The high displacement rates and synchronous movement along the basal and roof thrusts suggest that the ophiolite may have behaved as a microplate situated between western North America and an active arc from ~155 to 150 Ma. A steep thermal gradient was present in the Josephine-Galice thrust sheet from ~155 to 150 Ma, with amphibolite facies conditions developed along the basal thrust. After accretion of the ophiolite by underthrusting, the ophiolite and overlying flysch underwent low-grade dynamothermal regional metamorphism from 150 to 135 Ma. The upper age limit is tightly constrained by a 135 Ma K-feldspar cooling age, syntectonic plutons as young as 139 Ma, and a Lower Cretaceous angular unconformity. Very rapid exhumation is indicated by the late Valanginian to Hauterivian age (~130 Ma) of the unconformably overlying strata, suggesting unroofing by extensional tectonics.

INTRODUCTION

The two-fold division of ophiolites into Tethyan and Cordilleran types by *Moores* [1982] was a major step in realizing the diversity of ophiolites. The former are thrust over passive continental margins and have a metamorphic sole showing an inverted metamorphic gradient. Cordilleran ophiolites, on the other hand, form the basement for pelagic and volcaniclastic rocks in accreted tectonostratigraphic terranes. They are often incomplete and typically show evidence for having formed near active arcs, probably within back-arc or intra-arc basins [*Coleman*, 1984; *Saleeby*, 1990].

The Klamath Mountains of northwest California and southwest Oregon contain numerous Paleozoic and Mesozoic ophiolitic terranes which were progressively accreted by eastward underthrusting [Irwin, 1981; *Coleman*, 1984]. The Josephine ophiolite is the youngest and most complete of these accreted ophiolites and forms the basement for a Late Jurassic flysch sequence (Galice Formation). The Josephine has a full ophiolite sequence, but unlike Tethyan-type ophiolites, it is thrust beneath the continental margin. In turn, a magmatic-plutonic arc complex of similar age to the ophiolite is thrust beneath the Josephine. An amphibolitic sole occurs along this basal thrust, but there is no inverted metamorphic gradient [Harper et al., 1990]. The Josephine ophiolite is also different in that it is overlain by a thick flysch sequence and was regionally

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Paper number 93JB02061.
0148-0227/94/93JB-02061$05.00

metamorphosed and intruded by calc-alkaline plutons during emplacement. The Josephine ophiolite is well suited for a geochronologic study of its formation and thrust emplacement. Dating of ophiolite emplacement is greatly facilitated by an extensive suite of Late Jurassic calc-alkaline dikes and plutons. Regional and detailed mapping along with petrologic studies have established an extensive geologic framework for geochronologic studies [*Wells et al.*, 1949; *Dick*, 1976; *Snoke*, 1977; *Harper*, 1980, 1984; *Harper et al.*, 1990]. The geochronologic results are critical for constraining tectonic models and for correlation of extensional and orogenic events with changes in North American and Pacific plate motions [e.g., *Engebretson et al.*, 1985; *May et al.*, 1989].

In this paper we present U/Pb and $^{40}Ar/^{39}Ar$ geochronologic data that reveal a rapid sequence of events from spreading ridge petrogenesis to thrust emplacement of the ophiolite, followed by a long period of regional metamorphism and plutonism. One of the primary results of this study is that displacement on the roof and basal thrusts was rapid and synchronous (~155-150 Ma). Thrusting was followed by regional metamorphism during which the Galice flysch was penetratively deformed. Both the thrusting and subsequent regional metamorphism are part of the "Nevadan orogeny."

The Nevadan orogeny is a distinct Late Jurassic regional deformation event that profoundly influenced the tectonic and paleogeographic history of the southwestern U.S. Cordillera [*Blackwelder*, 1914; *Lanphere et al.*, 1968; *Burchfiel and Davis*, 1972, 1975, 1981; *Schweickert et al.*, 1984; *Harper and Wright*, 1984; *Wyld and Wright*, 1988; *Wright and Fahan*, 1988]. We follow *Bateman and Clark* [1974], *Schweickert et al.* [1984], *Harper and Wright* [1984], *Wyld and Wright* [1988], and *Wright and Fahan* [1988] in restricting the definition of the Nevadan orogeny to deformation and regional metamorphism that affected strata as young as Kimmeridgian. *Harper and Wright* [1984] estimated the Nevadan orogeny to have occurred between 150 and 145 Ma, which represents the age span for penetrative deformation of the Galice flysch. The geochronologic data presented in this paper, however, indicate that displacement on the roof and basal thrusts of the Josephine-Galice terrane occurred between ~155 and 150 Ma. During this time the Galice flysch was being deposited on the Josephine ophiolite, suggesting that it is a synorogenic deposit. We will also present evidence showing that metamorphism and at least local deformation continued until 135 Ma, at which time there was rapid exhumation followed closely in time by deposition of Lower Cretaceous strata. Continuation of the Nevadan orogeny until ~135 Ma is consistent with recent geochronologic studies from the Sierran foothills [*Saleeby et al.*, 1989*a*; *Tobisch et al.*, 1989].

GEOLOGIC SETTING

The Josephine ophiolite of northwestern California and southwestern Oregon is a large and intact ophiolite, having a basal peridotite covering >1000 km^2 (Figures 1 and 2). The crustal sequence of the ophiolite, including a thick sheeted dike complex, is exposed along strike for >130 km, and remnants of the ophiolite occur as much as 125 km to the south (Devils Elbow remnant, Figure 1 [*Wyld and Wright*, 1988]). Although the ophiolite formed in a suprasubduction zone environment [*Harper and Wright*, 1984; *Harper*, 1984, 1989; *Wyld and Wright*, 1988], it contains a complete crust-upper mantle sequence indicative of juvenile oceanic lithosphere formed by seafloor spreading. This study focuses on the most contiguous and complete exposures of the ophiolite in terms of internal stratigraphy and a complete structural succession from the basal thrust to the roof thrust. The Josephine ophiolite and Galice Formation are within the Smith River subterrane (Figures 2 and 3), which is bounded on the base and top by the Madstone Cabin and Orleans thrusts, respectively. The ophiolite is conformably overlain by ~50 m of radiolarian-rich hemipelagic rocks that grade rapidly upward into flysch of the Late Jurassic Galice Formation (Figure 4a) [*Harper*, 1983, 1984; *Pinto-Auso and Harper*, 1985; *Pessagno and Blome*, 1990]. Similar relationships are present in the Devils Elbow remnant of the Josephine ophiolite in the southern Klamath Mountains (Figure 1), except that the ophiolite is built on a basement of older Klamath rocks (Figure 4d) [*Wyld and Wright*, 1988]. Within the study area, the Galice locally appears to be underlain by older Klamath basement rather than the Josephine ophiolite, consisting of the Lems Ridge olistostrome and underlying pillow lavas (Figure 4c) [*Harper et al.*, 1985; *Ohr et al.*, 1986; *Ohr*, 1987]. An ophiolitic megabreccia at the top of the olistostrome is chemically identical to Josephine dikes and lavas [*Ohr*, 1987] and may be equivalent to ophiolitic breccias in the Devils Elbow remnant (Figure 4d). These local occurrences of older basement probably represent a rift edge facies formed during rifting and initial opening of the Josephine back-arc basin [*Wyld and Wright*, 1988].

The roof thrust is named the Preston Peak fault in the study area [*Snoke*, 1977] but is regionally known as the Orleans thrust [Jachens et al., 1986]. A large reentrant extending into the central Klamath Mountains (Figure 2) indicates >40 km of westward displacement on the roof thrust, and Jachens et al. [1986] used geophysical data to infer >110 km offset. The hanging wall of the Orleans thrust is the Late Triassic to Lower Jurassic Rattlesnake Creek terrane [Irwin, 1981], consisting of disrupted ophiolitic rocks overlain by stratified mafic volcanic and sedimentary rocks [*Gorman*, 1985; *Gray*, 1986; *Wyld and Wright*, 1988; *Saleeby*, 1990]. Also included with the Rattlesnake Creek terrane in Figure 1 are correlative high-grade metamorphic rocks of the Marble Mountains terrane [*Donato*, 1987; *Thompson et al.*, 1988]. Southeast of Gasquet, a complete assemblage characteristic of the Rattlesnake Creek Terrane is present in the upper plate of the Orleans fault [*Gorman*, 1985], but east of Gasquet, the upper plate consists of ultramafic rocks intruded and overlain by a mafic complex consisting of dikes and breccia (Preston Peak ophiolite of *Snoke* [1977]). U/Pb zircon dating suggests that the mafic complex is similar in age to the Josephine ophiolite [*Saleeby et al.*, 1982; *Saleeby and Harper*, in review], whereas the ultramafic basement is part of the Rattlesnake Creek terrane [*Gorman*, 1985]. The Preston Peak ophiolite has been interpreted as a rift edge facies formed as older Klamath basement was unroofed and intruded by mafic lavas during initial rifting of the Josephine basin [*Saleeby et al.*, 1982].

The base of the Josephine-Galice thrust sheet (Smith River subterrane) is exposed along its northern boundary where it structurally overlies intrusive rocks of the Chetco complex along the Madstone Cabin thrust (Figure 2). This basal thrust is associated with an amphibolitic sole up to 200 m thick (Figure 5) [*Dick*, 1976; *Cannat and Boudier*, 1985; *Harper et al.*, 1990; *Grady*, 1990]. The eastern extent of the Madstone Cabin thrust is uncertain, but it may continue eastward to where the Josephine ophiolite structurally overlies the Rogue Valley subterrane (Figure 2) [*Dick*, 1976; *Harding*, 1987].

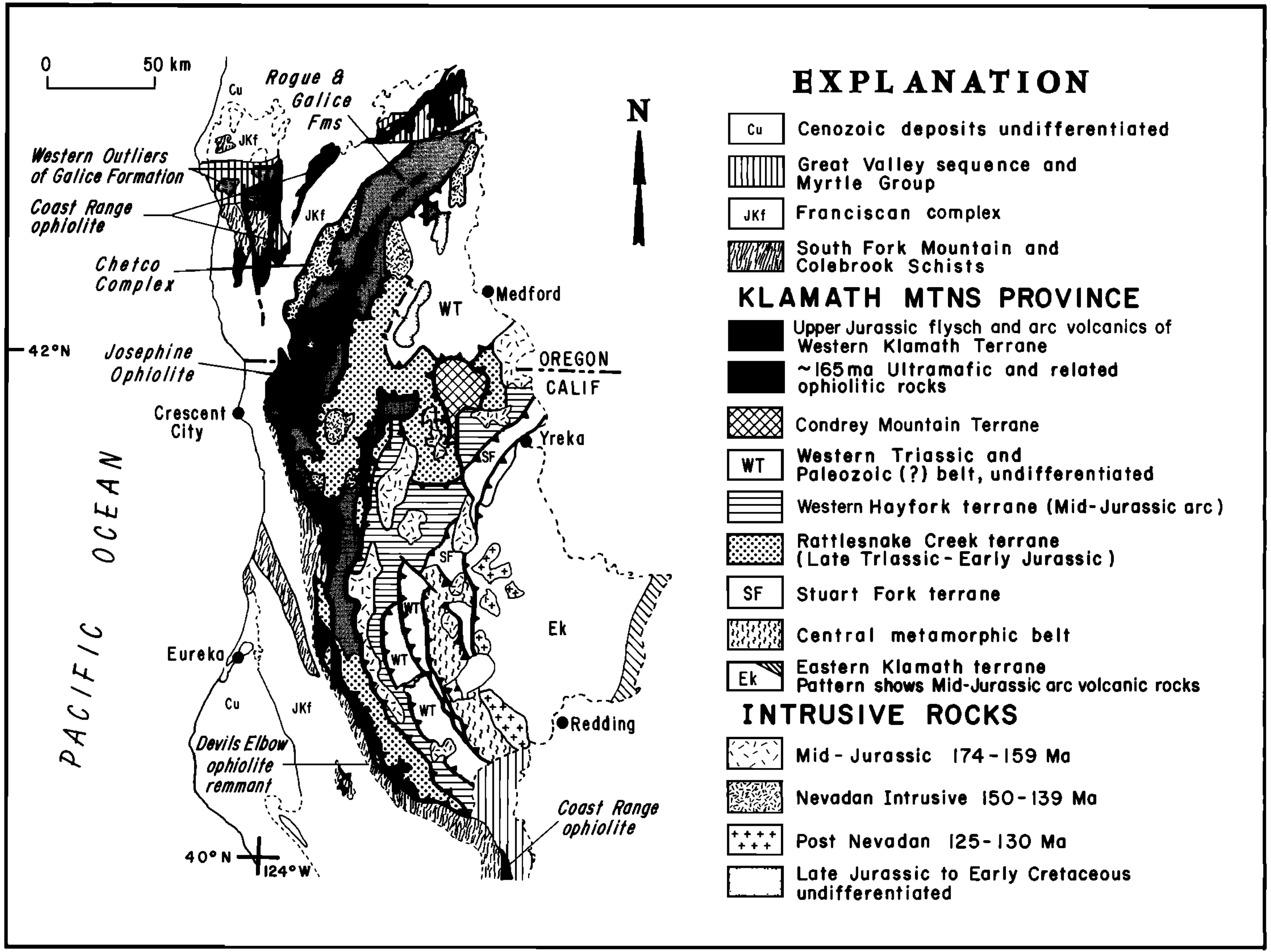


Fig. 1. Generalized geologic map of the Klamath Mountains, modified from *Irwin* [1981], *Coleman* [1972], *Page et al.* [1981] and *Blake et al.* [1985].

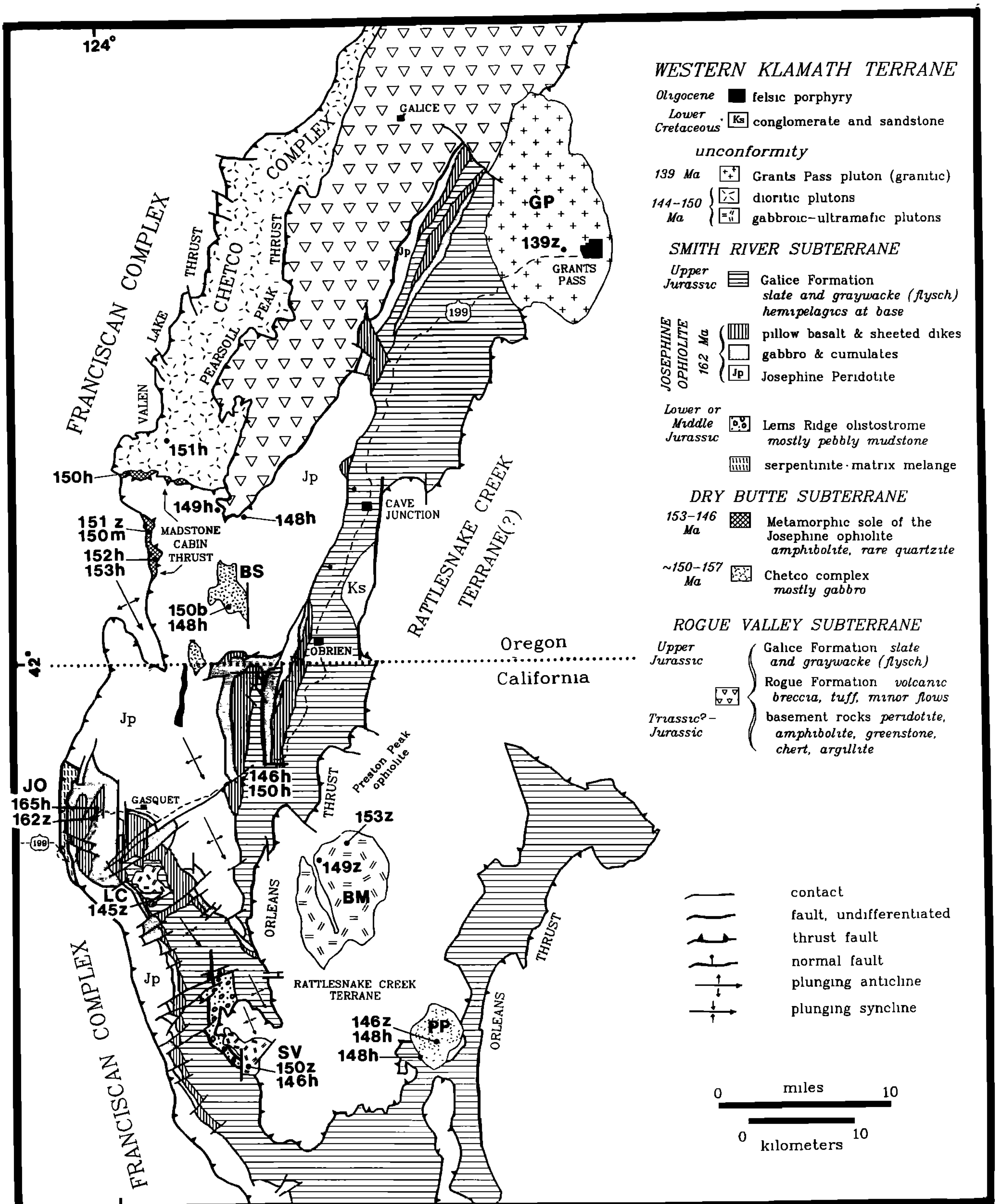


Fig. 2. Generalized geologic map of the west central Klamath Mountains, modified from *Harper* [1984], *Wagner and Saucedo* [1987] and, *Smith et al.* [1982]. Isotopic ages are from this study, except for the Bear Mountain (BM) complex [*Saleeby et al.*, 1982]. Abbreviations are b, biotite $^{40}Ar/^{39}Ar$; h, hornblende $^{40}Ar/^{39}Ar$; m, muscovite $^{40}Ar/^{39}Ar$; z, U/Pb zircon; BS, Buckskin Peak; GP, Grants Pass; JO, Josephine ophiolite; LC, Lower Coon Mountain; PP, Pony Peak; SV, Summit Valley.

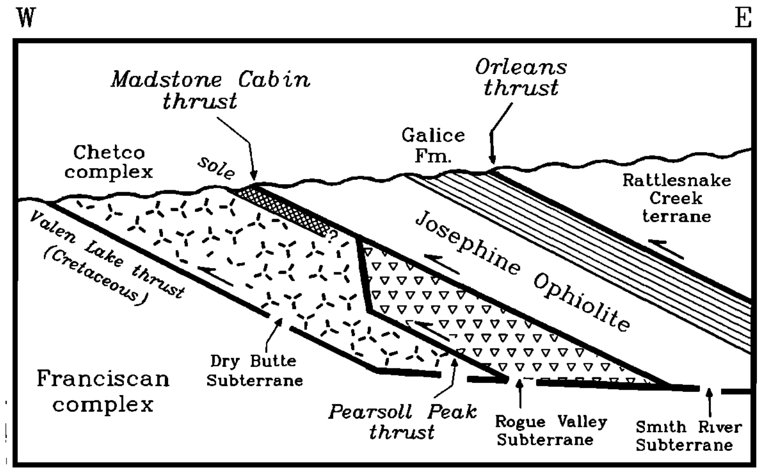


Fig. 3. Schematic cross section showing structural stacking of subterranes in the Western Klamath terrane.

The Rogue Valley subterrane includes the Rogue Formation, a thick sequence of mafic-to-silicic breccias, tuffs, and minor lava flows (Figure 4a) [*Garcia*, 1982; *Riley*, 1987]. The Rogue has yielded a 157±1.5 Ma U/Pb zircon age [*Saleeby*, 1984] and is conformably overlain by tuffaceous chert and flysch of the type Galice Formation [*Pessagno and Blome*, 1990; *Harper*, 1989]. The Galice flysch is virtually identical in lithology, petrography, chemistry, and age to the "Galice" flysch overlying the Josephine ophiolite (Figure 4b) [*Harper*, 1983, 1989; *Harper and Wright*, 1984; *Park-Jones*, 1988; *Pessagno and Blome*, 1990]. Beneath the Rogue Formation are ophiolitic breccias, siliceous argillite and chert, amphibolite, serpentinite, and peridotite which compose the basement of the Rogue Formation (Figure 4a) [*Yule et al.*, 1992]. These rocks, as well as similar rocks within the Smith River subterrane (Lems Ridge olistostrome, basement of Devils Elbow ophiolite remnant; Figures 4c and 4d), probably represent older Klamath basement (Rattlesnake Creek terrane) rifted away from North America during formation of the Josephine back-arc basin [*Wyld and Wright*, 1988; *Yule et al.*, 1992].

Both the Smith River (Josephine-Galice) and Rogue River subterranes were regionally metamorphosed to low grade and intruded by calc-alkaline magmas during the Nevadan orogeny [*Harper and Wright*, 1984]. The Galice Formation in both subterranes has a generally east dipping slaty cleavage, a gently plunging stretching lineation, and variably plunging fold axes [*Park-Jones*, 1988; *Riley*, 1987; Jones, 1988; *Harper*, 1980, 1989, 1992].

The Rogue Valley subterrane is thrust over a mafic batholith (Chetco complex) along the Pearsoll Peak thrust (Figures 2 and 3). This thrust has received little study, but the following evidence suggests it is a continuation or splay of the Madstone Cabin thrust: (1) the Chetco complex forms the footwall of both thrusts, (2) rocks beneath both thrusts were strongly deformed under amphibolite facies conditions and have a north-northeast mineral lineation [*Ramp*, 1975; *Dick*, 1976; *Loney and Himmelberg*, 1977; *Yule and Saleeby, 1993*], and (3) muscovite from deformed granitic rocks beneath both thrusts have yielded ages of ~150 Ma [*Hotz*, 1971; *Dick*, 1976; *Ramp*, 1984; this study].

The Chetco complex (Dry Butte subterrane) is a batholith consisting predominantly of hornblende gabbrobut ranges from gabbro through quartz diorite [*Wells et al.*, 1949; *Dick*, 1976]. It is similar in age to the Rogue Formation and probably represents the plutonic roots of the Rogue Formation [*Dick*, 1976; *Harper and Wright*, 1984].

The western boundary of the Western Klamath terrane is a Cretaceous thrust fault (Valen Lake thrust/South Fork fault) which truncates the Madstone Cabin thrust (Figure 2). It places the Josephine ophiolite and Chetco complex over latest Jurassic to Lower Cretaceous metasedimentary rocks of the Franciscan complex [*Blake et al.*, 1985].

TECTONIC SETTING

Tectonic models for the generation and emplacement of the Josephine ophiolite are well constrained by stratigraphic, structural, geochemical, and geochronologic data [*Saleeby et al.*, 1982; *Harper*, 1984; *Harper and Wright*, 1984; *Harper et al.*, 1985; *Wyld and Wright*, 1988; *Wright and Fahan*, 1988; *Saleeby*, 1990; this study]. A suprasubduction zone setting for generation of the Josephine ophiolite is indicated by arc detritus in overlying sediments, regional geologic setting, and magmatic affinities to arc tholeiites [*Saleeby et al.*, 1982; *Harper*, 1984, 1989; *Harper and Wright*, 1984; *Pinto-Auso and Harper*, 1985; *Wyld and Wright*, 1988]. Although details of published models vary, the essential elements are the following [*Saleeby et al.*, 1982; *Harper and Wright*, 1984; *Wright and Fahan*, 1988; *Wyld and Wright*, 1988]: (1) a Middle Jurassic arc (Western Hayfork terrane and related plutons) was built along western North America and underwent an orogenic phase between 169 and 161 Ma; (2) rifting of the arc followed by seafloor spreading resulted in a back-arc basin (Josephine ophiolite) and migration of the active arc westward (Rogue volcanics and Chetco complex), thus forming an arc/back-arc basin/remnant arc triad; and (3) during the Late Jurassic Nevadan orogeny the arc, back-arc basin, and remnant arc were imbricated by thrusting. As discussed by *Harper et al.* [1985], *Wyld and Wright* [1988], and *Saleeby* [1990], it is likely that rifting and seafloor spreading occurred along an intra-arc wrench fault similar to the modern Andaman Sea.

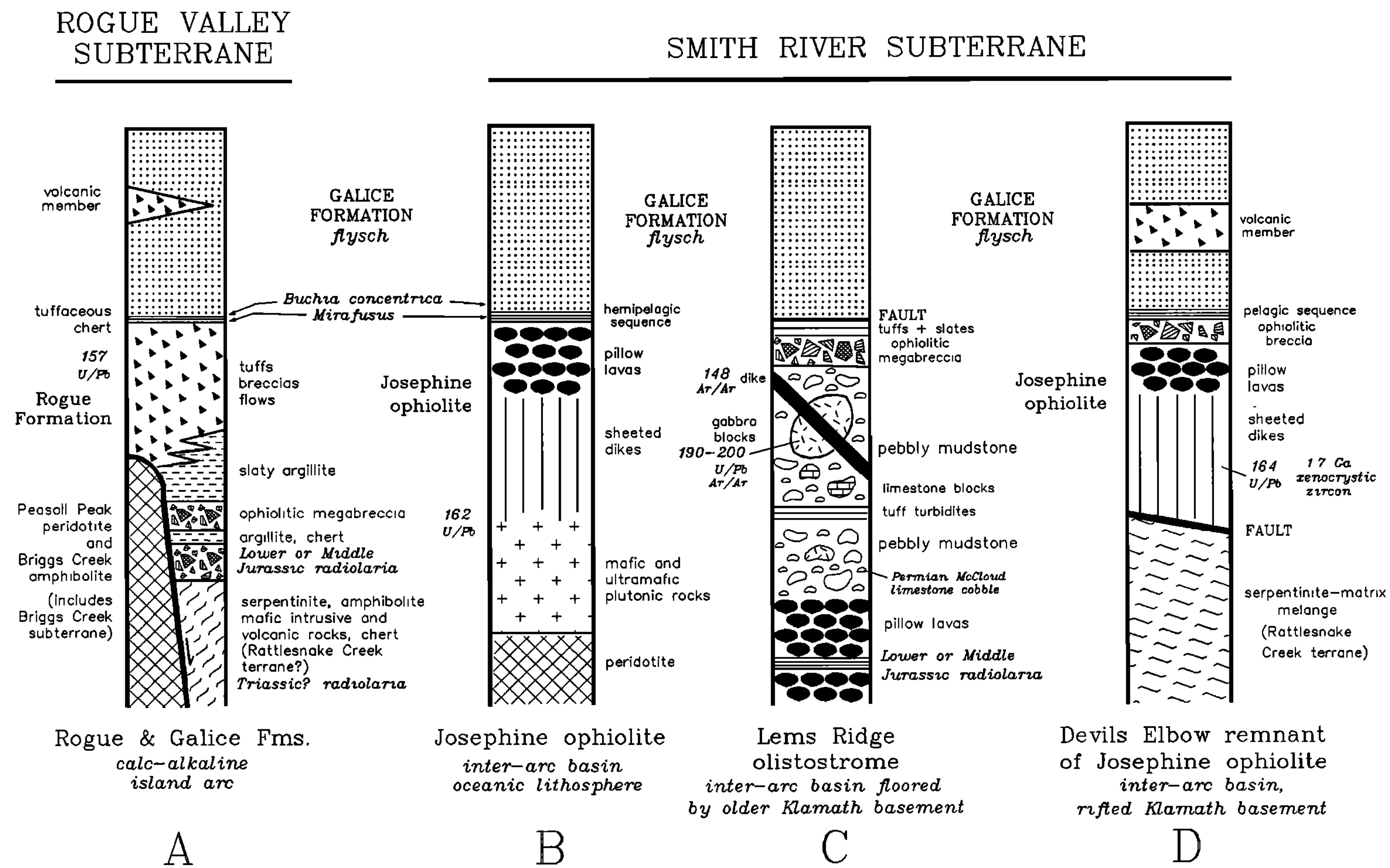


Fig. 4. (a) Tectonostratigraphy of the Rogue River subterrane structurally beneath the Josephine ophiolite [*Wells and Walker*, 1953; *Garcia*, 1982; *Saleeby*, 1984; *Riley*, 1987; *Yule and Saleeby*, 1993]. (b) Tectonostratigraphy of the Smith River subterrane in the west central Klamath Mountains [*Harper*, 1984, 1989]. (c) Tectonostratigraphy of the Lems Ridge olistostrome [*Harper et al.*, 1985; *Ohr*, 1987; *Ohr et al.*, 1987]. (d) Tectonostratigraphy of the Devils Elbow remnant of the Josephine ophiolite, southern Klamath Mountains [*Wyld and Wright*, 1988]. Fossil occurrences are from *Pessagno and Blome* [1990].

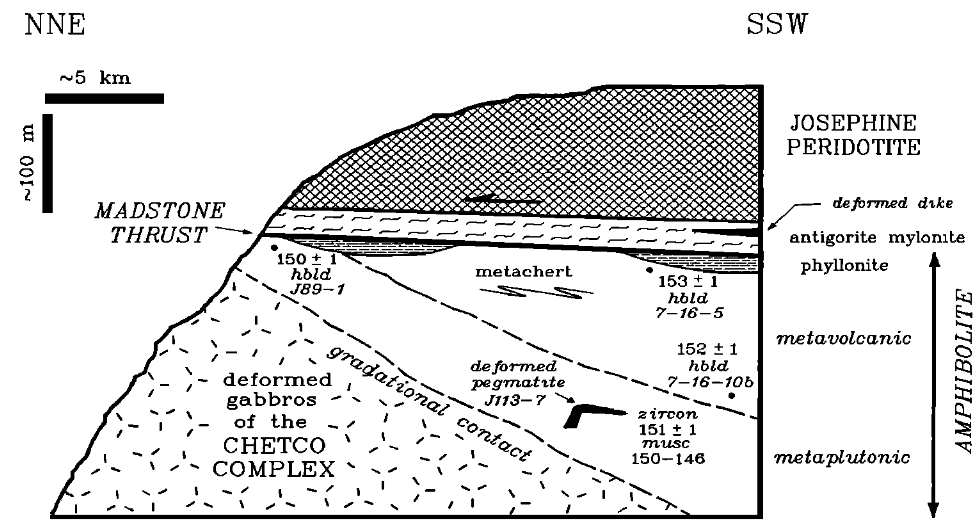


Fig. 5. Diagrammatic section through the amphibolite sole beneath the Josephine ophiolite modified from *Harper et al.* [1990] and *Grady* [1990].

METHODS AND ANALYTICAL PROCEDURES

Age determinations by the $^{40}Ar/^{39}Ar$ incremental heating method are listed in Table 1, and the step heating data are given in Appendix A (available on microfiche)[1]. Analytical procedures are summarized by *Harrison and Fitz Gerald* [1986], and all ages have been calculated with decay constants and isotopic abundances recommended by *Steiger and Jager* [1977].

Many $^{40}Ar/^{39}Ar$ ages were interpreted from plateaus defined by a number of successive similar incremental ages. The incremental ages are "apparent" ages because they are calculated under the assumption of an atmospheric composition for the trapped component of Ar ($^{40}Ar/^{36}Ar$ = 295.5). The presence of nonatmospheric components results in apparent ages that are too old as documented for several samples in this study. Even though nonatmospheric trapped Ar usually results in an irregular age spectrum, the sample age is often well constrained on an $^{36}Ar/^{40}Ar$ versus $^{39}Ar/^{40}Ar$ isochron diagram [*McDougall and Harrison*, 1988]. The isochron is a mixing line between trapped and radiogenic argon: the y intercept gives the composition of the trapped component ($^{36}Ar/^{40}Ar$), and the x intercept gives the radiogenic component ($^{39}Ar/^{40}Ar$) which is proportional to the age. Isochron ages are preferable, especially for low potassium minerals such as hornblende, because the composition of trapped Ar does not need to be assumed. Thus we use the isochron method whenever possible and have calculated ages and uncertainties using the *York* [1969] regression method. The data for some samples, however, form a cluster on the isochron diagram due to a constant radiogenic yield for each heating step. In these cases, there is no option but to assume atmospheric trapped Ar and calculate a plateau age by averaging the apparent ages.

Uncertainties reported for the $^{40}Ar/^{39}Ar$ ages in tables and shown on figures represent one standard deviation in analytical uncertainty. In addition, there is an estimated 1% uncertainty in age of the standard (flux monitor) analyzed with the unknowns. The combined uncertainty (analytical + flux monitor) is shown in parentheses in Table 1.

The $^{40}Ar/^{39}Ar$ plateau and isochron ages are cooling ages unless the mineral crystallized at temperatures below its closure temperature. Closure temperatures for amphiboles and micas are assumed to be ~500°C for igneous hornblende, ~350°C for muscovite, and ~300°C for biotite. For metamorphic hornblendes, especially those which exhibit exsolution, the closure temperature may be significantly lower than 500°C [*Harrison and Fitz Gerald*, 1986]. For example, a closure temperature of 440°C was experimentally determined by *Baldwin* [1988] for a metamorphic hornblende from Vermont. Closure temperatures for K-feldspar are variable but can be calculated from the step heating data.

The $^{40}Ar/^{39}Ar$ ages on igneous minerals will be similar to crystallization ages only if cooling was rapid (e.g., shallow level dikes). Many of the dated plutons are small and were intruded into rocks having ambient temperatures of $\lesssim$350°C. Such small intrusions would be expected to cool quickly below the ~500°C closure temperature for hornblende. Many of these plutons, however, consist of two or more phases, so the temperature may have remained elevated by continual addition of heat from new intrusions, especially if substantial volumes of magma were transported through the plutonic complex.

Zircon U/Pb age data are presented in Table 2, and the locations of the samples are given in Appendix A (available on microfiche). Notes on analytical procedures and assessment of uncertainties are given in the footnotes of Table 2. A more in-depth discussion of procedures and uncertainties is given by *Saleeby et al.* [1989b]. Discussion of the zircon data centers on $^{207}Pb/^{206}U$ versus $^{238}U/^{206}Pb$ concordia plots [*Tera and Wasserburg*, 1972]. For samples that are concordant, or very nearly concordant, and are of a restricted age range, the graphic relations are shown in a plot where equivalent segments of concordia are stacked with a common abscissa (Figure 6). This plot facilitates close examination of relations between concordia and error bars, which is particularly useful for geologically related samples. In the discussion of the zircon data, internal concordance is defined as agreement of the U/Pb and Pb/Pb ages of a given analysis within analytical uncertainty. It is expressed graphically by the intersection of the error bars with concordia.

[1] Appendix Tables A1 and A2 are available with entire article on microfiche. Order from American Geophysical Union, 2000 Florida Ave., N.W., Washington DC, 20009. Document B93-008; $2.50. Payment must accompany order.

TABLE 1. Summary of $^{40}Ar/^{39}Ar$ Ages

	Sample	Location	Description	Mineral	Plateau Age[a], Ma	Isochron Age[a,b], Ma	$^{40}Ar/^{36}Ar$ of Trapped Ar	Previous K/Ar Age[c]
			Josephine Ophiolite					
Smith River area	KH-6	124°01'36"W 41°59'45"N	Gabbro intruded by diabase dikes	hornblende	165.3 ± 3.1 (3.5)			
Devils Elbow remnant	OB-1	123°15'18"W 40°18'47"N	Plagiogranite clast in ophiolite breccia	hornblende		160.5 ± 2.5 (3.0)	282 ± 8	
	OB-4	same as OB-1	Gabbro clast in ophiolite breccia	hornblende		164.5 ± 5.0 (5.2)	377 ± 24	
			Amphibolite Sole of Josephine Ophiolite					
	7-16-10b	123°57'33"W 42°08'33"N	Amphibolite	hornblende	151.9 ± 1.0 (1.8)			
	7-15-2k	123°56'56"W 42°08'54"N	Amphibolite	hornblende	152.8 ± 0.8 (1.7)			
	J-113-7	123°58'00"W 42°09'18"N	Deformed pegmatite intruding amphibolite	muscovite	149.9 ± 0.3 (1.5)			147 ± 3
			Chetco Complex					
	J-89-1	123°58'54"W 42°11'54"N	Gneissic metagabbro	hornblende	150.5 ± 1.0 (1.8)			153 ± 3
	LCHB	123°55'42"W 42°12'26"N	Hornblendite	hornblende		151.4 ± 4.1 (4.4)	301 ± 51	155 ± 3
			Intrusives Cutting the Josephine Ophiolite and Galice Formation					
	J-97-6	123°49'38"W 42°09'24"N	Mafic dike intruding Josephine peridotite	hornblende	148.0 ± 2.6 (3.0)			150 ± 3
	J-98-12	123°51'36"W 42°09'54"N	Mafic dike intruding Josephine peridotite	hornblende	148.9 ± 2.1 (2.6)			146 ± 3
Buckskin Peak pluton	66	123°50'30"W 42°03'06"N	Quartz diorite	hornblende biotite[d]		148.6 ± 1.5 (2.1) 148.4 ± 1.2 (1.9) 149.7 ± 1.6 (2.2)	 1341 ± 110 725 ± 12	
	D24	123°48'50"W 41°52'06"N	Mafic sill in Galice Formation	hornblende	150.5 ± 1.4 (2.0)			
	D26	123°48'49"W 41°52'17"N	Mafic sill in Galice Formation	hornblende		146.2 ± 1.0 (1.8)	329 ± 16	
Lower Coon Mountain pluton	J84z	123°56'29"W 41°45'04"N	Granodiorite dike	K-feldspar		134.9 ± 0.9 (1.6)		
			Plutons Cutting the Orleans Thrust					
Summit Valley pluton	SV-1h	123°49'29"W 41°34'59"N	Hornblende gabbro	hornblende		144.1 ± 0.4 (1.5)	297 ± 7	
Pony Peak pluton	K85-26	123°33'25"W 41°34'36"N	Quartz diorite dike in contact aureole	hornblende		147.9 ± 0.1 (1.5)	298 ± 2	
	K85-53	123°32'38"W 41°35'40"N	Quartz diorite	hornblende		148.5 ± 2.0 (2.5)	476 ± 69	

[a] Errors are 1 standard deviation; errors in parentheses include uncertainty in J factor.
[b] Isochron ages and initial ratios are intercepts resultant from regression method of York [1969].
[c] K/Ar ages reported by Dick [1976] for same hand samples, recalculated using stadardized decay constants [Steiger and Jager, 1977].
[d] From Heizler and Harrison [1988]. See text for discussion of multiple isochron ages.

TABLE 2. Zircon Isotopic Age Data

Sample	Rock Unit	Location	Rock Type	Fraction[a], µm	Amount Analyzed, mg	Concentration, ppm: ^{238}U	Concentration, ppm: $^{206}Pb^{*\,c}$	Atomic Ratios: $^{206}Pb/^{204}Pb$	Atomic Ratios: $^{206}Pb^{*}/^{238}U$ c,d	Atomic Ratios: $^{207}Pb^{*}/^{235}U$ c	Atomic Ratios: $^{207}Pb^{*}/^{206}Pb^{*}$ c,d	Isotopic ages, Ma[b]: $^{206}Pb^{*}/^{238}U$	Isotopic ages, Ma[b]: $^{207}Pb^{*}/^{235}U$	Isotopic ages, Ma[b]: $^{207}Pb^{*}/^{206}Pb^{*}$
1	Grants Pass pluton	123°22'30"W 42°26'24"N	quartz diorite	<45	5.9	919	17.4	10681	0.02191(28)	0.1490	0.04936(18)	140	141	164±9
				45-62	7.0	744	14.2	4052	0.02214(19)	0.1524	0.04994(15)	141	144	192±7
				62-80	13.2	602	11.8	4229	0.02263(19)	0.1580	0.05065(13)	144	149	224±6
				80-100	11.2	385	7.5	1170	0.02246(18)	0.1584	0.05116(14)	143	149	248±7
2	Lower Coon Mtn pluton	123°56'12"W 41°45'15"N	quartz diorite	<45	6.6	467	9.3	3219	0.02308(16)	0.1565	0.04920(10)	147	148	157±6
				45-62	4.1	444	9.1	2439	0.02379(20)	0.1633	0.04979(12)	152	154	185±7
				62-80	4.8	246	5.0	2090	0.02348(20)	0.1601	0.04947(13)	150	151	170±7
				80-100	3.8	243	5.1	3255	0.02415(17)	0.1664	0.05001(10)	154	156	195±6
3	Pony Peak pluton	123°33'25"W 41°34'36"N	quartz diorite	<80c	~0.3	~410	~8	1011	0.02321(27)	0.1580	0.04940(28)	148	149	167±13
				<45	4.1	593	12.1	3463	0.02366(19)	0.1651	0.05065(13)	151	156	224±7
				45-80	7.0	265	5.5	2970	0.02397(19)	0.1693	0.05123(15)	153	157	250±8
				80-100	19.6	238	4.9	1975	0.02390(22)	0.1671	0.05075(18)	152	157	229±9
				80-120c	2.5	365	7.4	3549	0.02341(14)	0.1640	0.05082(24)	149	154	232±12
4	Summit Valley plutonic complex	123°49'29"W 41°34'59"N	hornblende gabbro	45-80	13.5	122	2.5	7373	0.02370(19)	0.1601	0.04902(12)	151	151	149±6
				80-120	19.0	80.5	1.64	7202	0.02359(19)	0.1595	0.04908(12)	150	150	151±6
5	amphibolite sole	123°58'00"W 42°09'18"N	granitic pegmatite	<62	3.1	546	11.2	3966	0.02370(15)	0.1603	0.04909(07)	151	151	152±4
6	Josephine ophiolite	124°02'05"W 41°50'16"N	plagiogranite	<80a	2.1	379	8.4	1919	0.02547(19)	0.1731	0.04932(10)	162	162	163±5

[a] Fractions separated by grain size and magnetic properties. Magnetic properties are given as nonmagnetic split at side/front slopes for 1.7 amps on Franz Isodynamic Separator. Samples hand-picked to 99.9% purity prior to dissolution; c, clear perfectly euhedral grains only; a, grains abraded by technique similar to ***Krogh*** [1982]. Dissolution and chemical extraction techniques modified from ***Krogh*** [1973].

[b] Decay constants used in age calculation: $\lambda^{238}U = 1.55125 \times 10^{-10}$, $\lambda^{235}U = 9.8485 \times 10^{-10}$ [***Jafey et al.***, **1971**] $^{238}U/^{235}U$ atomic = 137.88. Uncertainties calculated by quadratic sum of total derivatives of ^{238}U and $^{206}Pb^{*}$ concentration and $^{207}Pb^{*}/^{206}Pb^{*}$ equations with error differentials defined as (1) isotopic ratio determinations from standard errors ($\sigma/\sqrt{n}$) of mass spectrometer runs plus uncertainties in fractionation corrections based on multiple runs of NBS 981, 982, 983, and U500 standards; (2) spike concentrations from range of deviations in multiple calibrations with normal solutions; (3) spike compositions from external precisions of multiple isotope ratio determinations; (4) uncertainty in natural $^{238}U/^{235}U$ from ***Chen and Wasserburg*** [1981]; and (5) nonradiogenic Pb isotopic compositions from uncertainties in isotope ratio determinations of blank Pb and uncertainties in composition of initial Pb from references given above.

[c] Radiogenic-nonradiogenic correction based on 40 picogram blank Pb (1:18.78:15.61:38.50) and initial Pb approximations: 1:18.6:15.6:38.2 [***Stacy and Kramers***, **1975**] and 1:18.21:15.51:37.86 for sample 5 [**after *Chen and Shaw*, 1982.**]

[d] Uncertainties in $^{206}Pb^{*}/^{238}U$ and $^{207}Pb^{*}/^{206}Pb^{*}$ given as (±) of last two values.

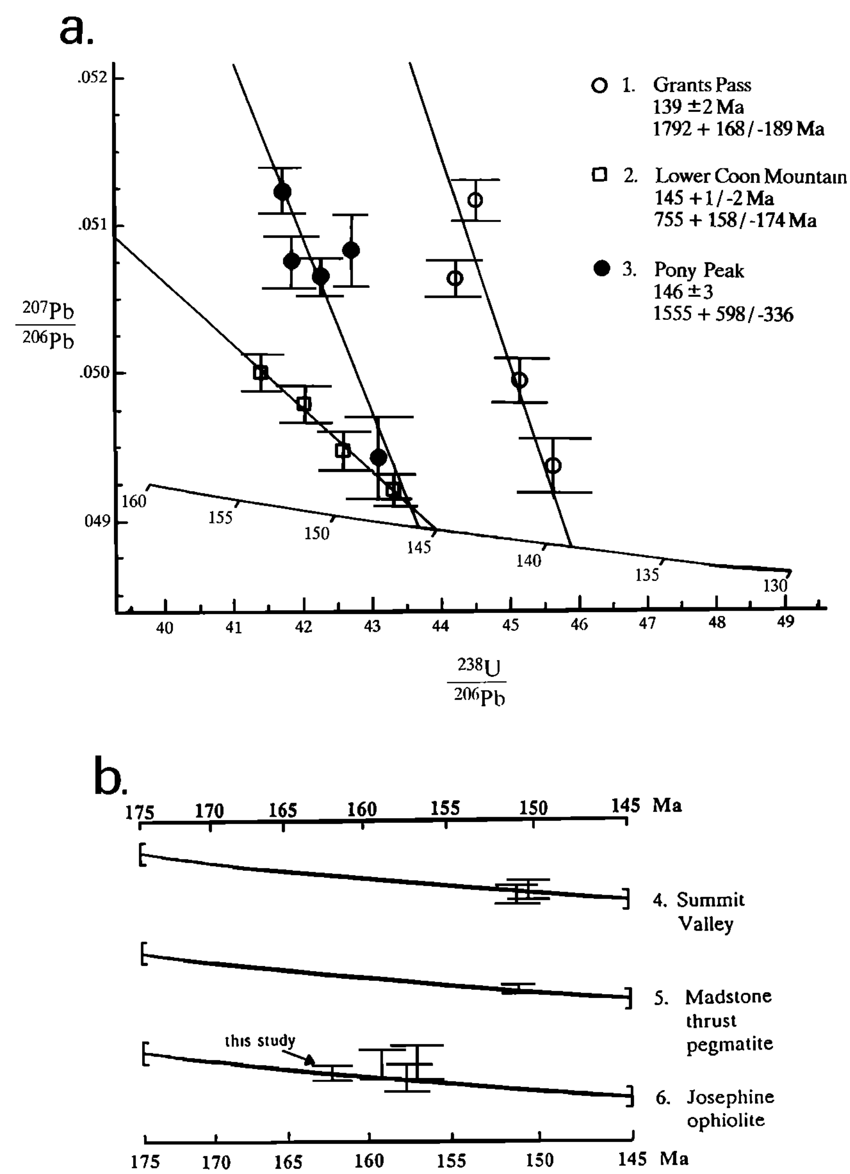


Fig. 6 (a) $^{207}Pb/^{206}Pb$ - $^{238}U/^{206}Pb$ concordia diagram [*Tera and Wasserburg*, 1972] for sample 1 data from Grants Pass pluton, sample 2 data from the Lower Coon Mountain pluton, and sample 3 data from Pony Peak pluton. Linear regression and intercept errors adapted from *York* [1969]. (b) Diagram showing stacked segments of concordia from $^{207}Pb/^{206}Pb$-$^{238}U/^{206}Pb$ concordia plots arranged for close comparison of error-bar-concordia relations for concordant and slightly discordant samples. Bars at ends of concordia segments show uncertainties in values of $^{207}Pb/^{206}Pb$ for concordia resulting from uncertainties in ^{238}U and ^{235}U decay constants, based on *Mattinson's* [1987] treatment of data from *Jaffey et al.* [1971].

External concordance is defined as agreement, or graphic overlap, in internally concordant ages from multiple fractions of a given sample or from multiple samples collected from the same rock unit. Externally concordant ages are considered to be accurate approximations of igneous ages in as much as the effects of multi-stage isotopic evolution have been observed to vary with the physical properties of zircon and are also heterogeneous within map-scale units. Treatment of discordant samples requires the application of a discordance mechanism model which is best developed by assessment of the isotopic systematics in the geological context of the sample.

THE JOSEPHINE OPHIOLITE

Saleeby et al. [1982] reported two 157±2 Ma U/Pb zircon ages on plagiogranites located 4 to 5 km west of Gasquet, and *Wright and Wyld* [1986] reported a 164±1 Ma U/Pb zircon age from the Devils Elbow remnant of the ophiolite exposed in the southern Klamath Mountains (Figure 4d). Radiolaria faunas from mudstones interbedded with pillow lavas of the ophiolite indicate a late Callovian age [*Pessagno and Blome*, 1990].

Data points for the previous zircon analyses from the two plagiogranites near Gasquet are shown on Figure 6b (sample 6, note large error bars). The remaining zircon population from one of the plagiogranites (A88z) was treated by air abrasion to a well-rounded state which, along with improved analytical procedures and common Pb corrections, yields an upward shift along concordia and a significantly smaller error bar. The resulting internally concordant age is 162±1 Ma, which is taken to be a better approximation of the igneous age for the plagiogranite than ages derived from the previous analyses. Air abrasion removal of the rims resulted in a significant decrease in U concentration (compare data in Table 2 with Table 1 of

Saleeby et al. [1982]) indicating very high U concentrations in the rims. Small amounts of Pb loss or U gain in grain rims are likely to have induced the dispersion of data points in the earlier analyses (Figure 6b).

Hornblende from a high-level gabbro (KH-6h) was dated by $^{40}Ar/^{39}Ar$ (Table 1). The sample locality is 4 km west of Gasquet (Figure 2) and 1 km north of the dated plagiogranite (A88z). A wide variety of amphibole compositions and textures is present which formed during retrogressive subseafloor hydrothermal metamorphism [*Harper et al.*, 1988; Kimball, 1988]. Most of the hornblende is deep-brown (deuteric?) and high in Al, and it has 0.23 wt % K_2O and K/Ca=0.023. The sample also contains some low-Al green hornblende (0.16 wt % K_2O, K/Ca=0.011) and actinolite [Kimball, 1988]. The $^{40}Ar/^{39}Ar$ analysis gave a complex release spectrum. Most of the steps have a K/Ca ratio similar to the brown hornblende (Figure 7a), but the last step has a K/Ca ratio closer to the green hornblende. The last four steps, comprising >45% of the total ^{39}Ar released, yields a plateau age of 165±3 Ma, concordant with the 162±1 Ma zircon age in the nearby plagiogranite. This age is interpreted as an igneous age because high-level gabbros were cooled below 500°C before the end of dike injection (i.e., on axis) [*Alexander and Harper*, 1992].

Samples OB-1 and OB-4 are plutonic clasts within an ophiolitic breccia overlying pillow basalts of the Devils Elbow ophiolite remnant of the Josephine ophiolite (Figure 4d). The breccia is apparently a talus deposit derived from the underlying ophiolite [*Wyld and Wright*, 1988]. The two samples are clasts containing hydrothermal hornblende, and retrogressive metamorphism is evident from the presence of albite, chlorite, epidote, prehnite, and pumpellyite. Sample OB-1 is a plagiogranite containing 15% quartz and relict clinopyroxene largely replaced by yellow-brown hornblende, whereas sample OB-4 is a metagabbro having mottled brown and green hornblende. Both samples display complex age spectra which are the result of nonatmospheric trapped Ar as evident from the isochron diagrams (Figure 7b and 7c). The samples give linear arrays on the isochron diagram, yielding isochron ages of 161±3 and 165±5 Ma. The occurrence of these samples as clasts in an ophiolitic breccia indicates rapid uplift and exposure on the seafloor [*Wyld and Wright*, 1988]. Thus these $^{40}Ar/^{39}Ar$ cooling ages are interpreted to be equivalent to igneous ages within the resolution of the dating, consistent with the 164±1 Ma U/Pb zircon age for nearby plagiogranite [*Wright and Wyld*, 1986].

In summary, the revised 162±1 Ma zircon age and the 164±1 zircon age for the Devils Elbow remnant [*Wright and Wyld*, 1986] are considered to be close estimates of igneous ages for the Josephine ophiolite in the these two areas. The three $^{40}Ar/^{39}Ar$ hornblende ages of 165±3, 161±3, and 165±5 Ma are concordant with the zircon ages, but they have greater analytical uncertainty because of complex age spectra and low K_2O contents.

BASAL THRUST AND AMPHIBOLITE SOLE

Samples from the amphibolite sole were dated in order to constrain the age of penetrative deformation associated with the Madstone Cabin thrust. Field and petrologic studies have shown that there is no inverted metamorphic gradient [*Harper et al.*, 1990]. Instead, both the amphibolite and underlying mafic batholith (Chetco complex) underwent retrogressive metamorphism from amphibolite to greenschist facies [*Wells et al.*, 1949; *Dick*, 1976; *Loney and Himmelberg*, 1977; *Harper et al.*, 1990; *Grady*, 1990]. In order to constrain this cooling history and timing of ductile thrusting we obtained $^{40}Ar/^{39}Ar$ age spectra for hornblende and muscovite and a zircon age for a syntectonic pegmatite. Some of the samples were previously dated by the K/Ar technique [*Dick*, 1976].

Amphibolite Sole

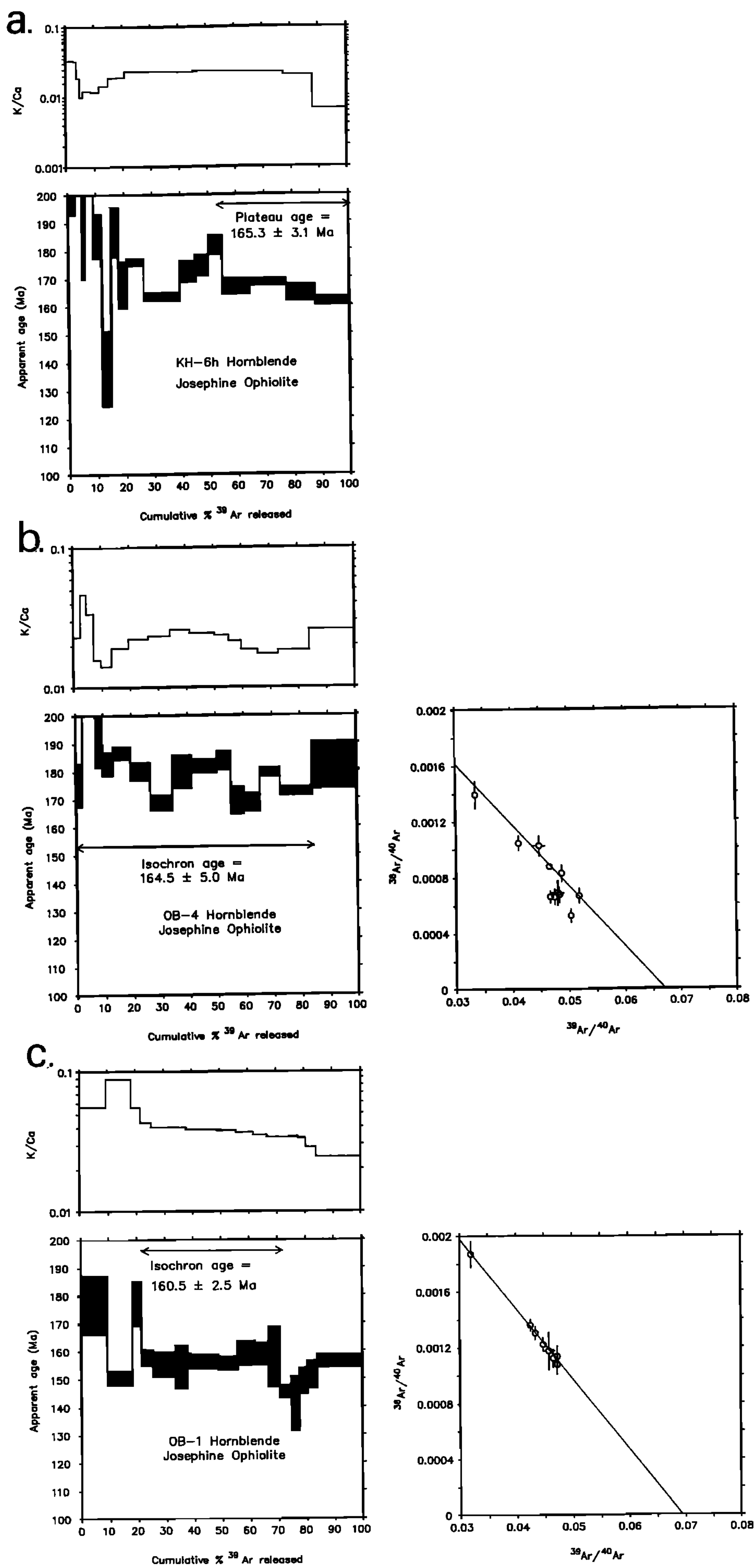
As much as 200 m of gneissic amphibolite occurs directly beneath the Madstone Cabin thrust (Figure 5) and has been interpreted as a synthrusting metamorphic sole [*Armstrong and Dick*, 1974; *Dick*, 1976; *Cannat and Boudier*, 1985; *Harper et al.*, 1990; *Grady*, 1990]. A diagrammatic north-south cross section of the Madstone Cabin thrust and amphibolite sole is shown in Figure 5. The upper plate of the thrust consists of remarkably fresh, unserpentinized peridotite of the Josephine ophiolite. Along the thrust, the peridotite has been converted to a 20-40 m thick serpentinite mylonite formed at ~500°C [*Grady*, 1990; *Harper et al.*, 1990]. A few meters of phyllonites occur locally directly beneath the serpentinite mylonite (Figure 5) and formed by retrogressive greenschist-facies metamorphism of amphibolite [*Harper et al.*, 1990]. The serpentinite mylonite, phyllonite, and amphibolite all have mylonitic fabrics and a north-northeast trending lineation. Shear indicators indicate north-northeast directed thrusting of the ophiolite over the Chetco complex [*Cannat and Boudier*, 1985; *Grady*, 1990; *Harper et al.*, 1990].

Much of the amphibolite sole beneath the Madstone Cabin thrust appears to have been derived from the underlying Chetco complex: (1) Chetco gabbros grade upward into the amphibolitic sole with increasing deformation [*Wells et al.*, 1949; *Ramp*, 1975; *Dick*, 1976; *Loney and Himmelberg*, 1977], and (2) amphibolites in the lower part of the sole generally have plagioclase porphyroclasts suggestive of a gabbro protolith and chemical affinities to the Chetco complex [*Grady*, 1990]. Along the southern strand of the thrust (Chetco Lake area), amphibolite in the upper part of the sole appears to be metavolcanic in origin based on the presence of micaceous quartzite (metachert; Figure 5) and chemical affinities to mid-ocean ridge and within-plate basalts [*Grady*, 1990]. The protolith age and origin of the metavolcanic rocks are unknown, and derivation from the lavas of the Josephine ophiolite is unlikely because of very different chemical affinities [*Harper*, 1984, 1989]. They could, however, be wall rocks of the Chetco complex that were deformed along with the Chetco complex during thrusting,

Amphibolite samples 7-16-10b and 7-16-5 are from the upper metavolcanic part of the amphibolite sole. The $^{40}Ar/^{39}Ar$ analysis of hornblendes from these samples resulted in similar, complex release spectra (Figure 8a and 8b). Plateaus comprise ~70% of the total ^{39}Ar released and yield ages of 151.9±1.1 and 152.8±0.8 Ma, respectively. The final 30% of ^{39}Ar released is first characterized by an increase in age to 158 Ma, which then decreases back to 154 Ma. This pattern could reflect an incompletely outgassed protolith, slow cooling of a ~160 Ma amphibole, or a trapped nonatmospheric argon component which is not evident on an isochron diagram due to scatter in the data. Regardless of this complication, the 152 and 153 Ma plateau ages of these two samples are interpreted to date the time of cooling below their closure temperature. These ages are concordant with a 151±3 Ma K/Ar hornblende age obtained on a nearby amphibolite [*Dick*, 1976].

Syntectonic Pegmatite Cutting Amphibolite

Granitic pegmatite dikes containing igneous muscovite and garnet locally cut the amphibolite sole (Figure 5) [*Ramp*, 1975;



a.
K/Ca
0.1
0.01
0.001
Apparent age (Ma)
Plateau age = 165.3 ± 3.1 Ma
KH-6h Hornblende
Josephine Ophiolite
Cumulative % 39 Ar released
b.
Isochron age = 164.5 ± 5.0 Ma
OB-4 Hornblende
Josephine Ophiolite
36Ar/40Ar
39Ar/40Ar
c.
Isochron age = 160.5 ± 2.5 Ma
OB-1 Hornblende
Josephine Ophiolite

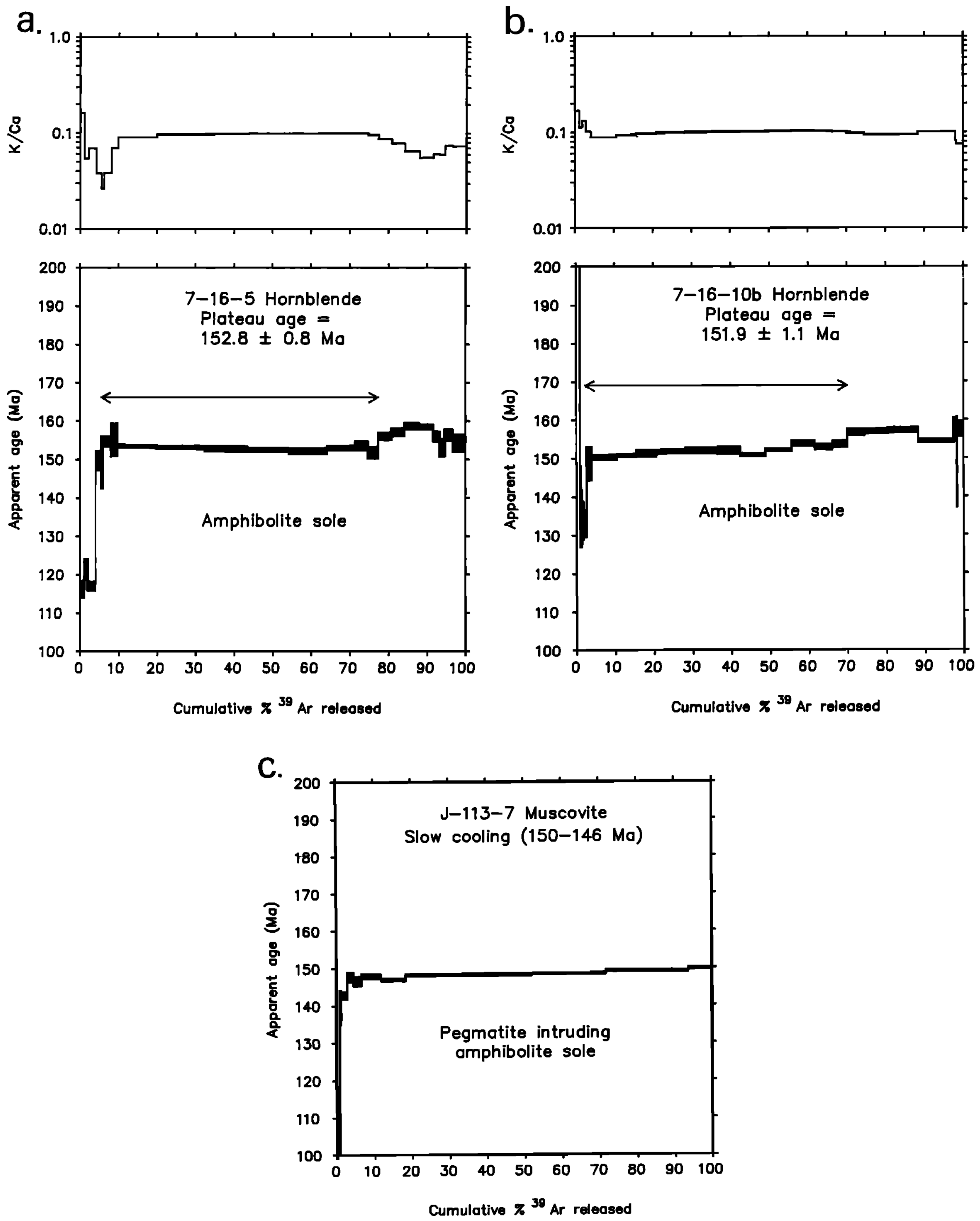


Fig. 8. (a) and (b) $^{40}Ar/^{39}Ar$ age spectra for hornblendes from two amphibolites of the amphibolite sole beneath the basal (Madstone Cabin) thrust (Figure 5). (c) $^{40}Ar/^{39}Ar$ age spectrum for muscovite from a deformed pegmatite dike cutting the amphibolite sole. U/Pb zircon data for this sample are shown in Figure 6b.

Fig. 7. (a) $^{40}Ar/^{39}Ar$ age spectrum for hornblende from a high-level gabbro in the Josephine ophiolite west of Gasquet. (b) and (c) $^{40}Ar/^{39}Ar$ age spectra for hornblende from two plutonic clasts in an ophiolitic talus breccia overlying the 164 Ma Devils Elbow remnant of the Josephine ophiolite [*Wyld and Wright*, 1988].

Dick, 1976]. They cut foliation in the amphibolite but are themselves penetratively deformed, indicating syntectonic intrusion [*Harper et al.*, 1990]. Prehnite is present as a metamorphic mineral, indicating that at least some recrystallization occurred at temperatures below amphibolite facies.

A ~25 kg sample of the pegmatite yielded ~3 mg of fine, uniform, well-faceted zircon that was free of inclusions or cores. Analysis of this zircon separate (sample 5, J-113-7) yielded an internally concordant age of 151±1 Ma (Table 2 and Figure 6b) and is considered to be an igneous age.

Deformed igneous muscovite from this dike exhibits a release spectrum having a linear age gradient ranging from 146 to 150 Ma (Figure 8c). The gradient is interpreted as slow cooling through the closure temperature for muscovite (~350°C).

Cooling Ages for the Chetco Complex

The Chetco complex is predominantly hornblende gabbro, much of which has been penetratively deformed under amphibolite-facies conditions [*Wells et al.*, 1949; *Ramp*, 1975; *Dick*, 1976; *Loney and Himmelberg*, 1977]. It consists of numerous intrusive phases, including large masses of late stage undeformed hornblendite and pegmatitic hornblende gabbro [*Dick*, 1976; *Loney and Himmelberg*, 1977]. Retrogressive greenschist-facies metamorphism is widespread [*Wells et al.*, 1949; *Dick*, 1976].

Hotz [1971b] and *Dick* [1976] reported six K/Ar ages for the Chetco complex which range from 153±4 to 160±3 Ma, including a concordant 154 Ma biotite-hornblende pair, and one age of 140±4 Ma. U/Pb zircon ages on the Chetco complex range from156 to 161 Ma [*Yule and Saleeby*, 1993].

Sample J-89-1 is a gneissic metagabbro located a few meters below the Madstone Cabin thrust along its northern strand [*Dick*, 1976]. Igneous hornblende and plagioclase are both largely recrystallized. In addition to strongly deformed metagabbro, outcrops at this locality include isoclinally folded amphibolite. Amphibolite at this locality has been mapped as either thin or absent [*Ramp*, 1975; *Dick*, 1976; *Loney and Himmelberg*, 1977; *Page et al.*, 1981]. Because the rocks at this locality are transitional between amphibolite and deformed metagabbro of the Chetco complex, it is somewhat arbitrary whether to assign them to the amphibolite sole or to deformed gabbro of the Chetco complex. In any event, the hornblende age for this sample records cooling below its closure temperature and is thus comparable to the two hornblende ages on amphibolite samples 7-16-10b and 7-16-5 (Figure 8). The $^{40}Ar/^{39}Ar$ analysis of hornblende from sample J-89-1 yielded a flat age spectrum for >75% of the ^{39}Ar released (Figure 9a). The plateau age is 150.3±1.0 Ma, concordant with a previous K/Ar age of 153±3 Ma for the same hornblende separate (Table 1) [*Dick*, 1976].

A hornblendite (LCHB, Table 1) was dated from an area of undeformed late stage pegmatitic hornblende gabbro located ~3 km north of the Madstone Cabin thrust. The $^{40}Ar/^{39}Ar$ analysis of hornblende resulted in an essentially flat age spectrum, and isochron analysis yields an age of 151±4 Ma (Figure 9b), concordant with a K/Ar hornblende age of 155±3 Ma for the same mineral separate (Table 1) [*Dick*, 1976]. This age records cooling below ~500°C and must be, at most, only a few million years younger than the igneous age since the hornblende pegmatite unit cuts older deformed phases of the Chetco complex dated at156 to 161 Ma (Pb/U zircon ages [*Yule and Saleeby*, 1993]).

Summary

The time of cooling below ~450-500°C for the amphibolite sole was ~151 Ma, based on the 152 and 153 Ma hornblende ages for the amphibolite, the 150 hornblende age for deformed metagabbro directly beneath the thrust, and the 151 Ma cooling age for late stage hornblendite in the underlying Chetco complex. The time of hornblende crystallization in the amphibolites was probably earlier than 151 Ma, since the closure temperatures for most hornblendes (~450-500°C) is below the lower limit of amphibolite facies (~550°C [*Liou et al.*, 1974] at estimated pressures of 2-4 kbar (200-400 MPa) [*Harper et al.*, 1990]). Cooling through the closure temperature of muscovite (~350°C) occurred from 150 to 146 Ma. The cooling history is consistent with the retrogressive metamorphism evident in the amphibolite sole and Chetco complex [*Harper et al.*, 1990].

The hornblende and muscovite ages imply that the 151 Ma pegmatite dike (zircon age) was intruded into amphibolite that was at temperatures near the amphibolite-to-greenschist facies transition. This conclusion is supported by the field and textural relationships suggesting syntectonic intrusion of the pegmatite.

The beginning of movement on the Madstone Cabin thrust is constrained to be younger than the ophiolite (~162 Ma) and younger than deformed intrusions of the Chetco complex (156 to 160 Ma [*Yule and Saleeby*, 1993]). Most of the displacement is inferred to have occurred under amphibolite-facies conditions because of the relatively small thickness of greenschist-facies phyllonite. Thus most of the displacement apparently occurred prior to ~151 Ma (i.e., at temperatures higher than the closure temperature for hornblende). Subsequent minor displacement under greenschist-facies conditions is inferred to have occurred from ~150 to 146 Ma as recorded by slow cooling of muscovite, although some deformation could have continued after cooling below ~350°C at 146 Ma. Thus the age range for movement on the Madstone Cabin thrust is inferred to be ~155 to ~145 Ma, with nearly all displacement completed by 150 Ma.

The field and petrographic relationships of the 151 Ma pegmatite shows that magmas were being intruded during thrusting [*Harper et al.*, 1990]. *Loney and Himmelberg* [1977] describe similar syntectonic gabbro dikes that cut amphibolite but are themselves deformed. These intrusions, and possibly the late stage hornblendite (151 Ma cooling age) and related hornblende gabbro pegmatite in the Chetco complex [*Dick*, 1976; *Loney and Himmelberg*, 1977], are probably related to abundant ~145-150 Ma dikes and plutons that cut the overlying Josephine ophiolite, Galice Formation, and Rogue Valley subterrane [*Dick*, 1976; *Saleeby et al.*, 1982; *Saleeby*, 1984; *Harper and Wright*, 1984; *Wyld and Wright*, 1988; this study].

AGE OF INTRUSIONS CUTTING THE JOSEPHINE OPHIOLITE AND GALICE FORMATION

Calc-alkaline dikes, sills, and plutons are common in the Josephine ophiolite and overlying Galice Formation. Most of the dikes and plutons are mafic in composition, but they range from mafic to granitic. These intrusions are abundant east of the longitude of Gasquet, but are rare west of this longitude [*Dick*, 1976; *Harper*, 1984]. The calc-alkaline affinity of the intrusions is apparent from the abundance of hydrous phases, porphyritic textures, and geochemistry [*Dick*, 1976; *Harper*, 1980, 1985; *Norman*, 1984]. The dikes and sills have clearly shared in the Nevadan regional metamorphism, and many are also deformed [*Harper*, 1984, 1992; *Wyld and Wright*, 1988].

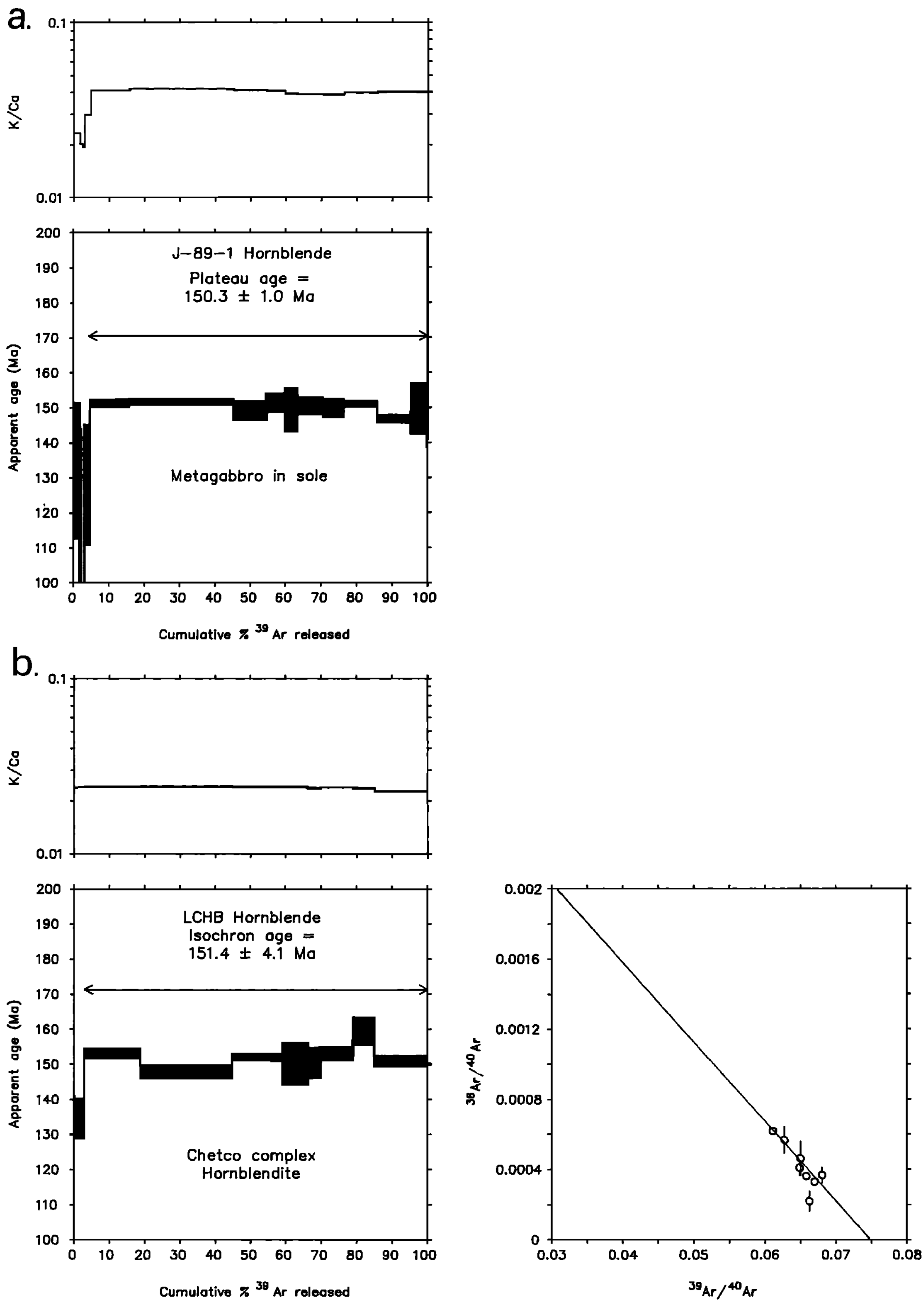


Fig. 9. (a) ^{40}Ar/^{39}Ar age spectrum for hornblende from a deformed amphibolite-facies metagabbro associated with amphibolite, located directly beneath the basal (Madstone Cabin) thrust. (b) ^{40}Ar/^{39}Ar age spectrum for a hornblendite collected from a late stage gabbro pegmatite and hornblendite unit in the Chetco complex.

Intrusions Cutting the Josephine Ophiolite

Dick [1976] obtained a number of ~145 to 155 Ma K/Ar ages for dikes and small plutons cutting the basal peridotite of the Josephine ophiolite. We analyzed two of Dick's hornblende separates (J-97-6, J-98-12) from dikes collected along the northern margin of the ophiolite, 12 and 14 km west of Cave Junction (Figure 2). The hornblendes yielded $^{40}Ar/^{39}Ar$ age spectra having similar shapes: initial ages of ~100-120 Ma rise to well defined plateaus of 148.0±2.6 and 148.9±2.1 Ma (Figure 10a and 10b). The plateau ages are interpreted as igneous ages for the dikes and are concordant with previous K/Ar ages of 150±3 and 146±3 Ma for these samples (Table 1) [*Dick*, 1976].

The Buckskin Peak pluton intrudes the Josephine peridotite just north of the California-Oregon border (Figure 2). It consists of hornblende diorite and biotite-hornblende quartz diorite [*Wells et al.*, 1949; *Harding*, 1987] and has not been studied in detail. Hornblende and biotite from a quartz diorite were dated (sample 66). The hornblende yielded a plateau age of 148.6±1.5 Ma (Figure 10c). The $^{40}Ar/^{39}Ar$ data for biotite from this sample have been reported by *Heizler and Harrison* [1988], who documented two distinct trapped Ar components and derived ages of 149.7±1.6 and 148.4±1.2 Ma (Table 1) using isochron analysis. The concordant hornblende (149 Ma) and biotite (150, 148 Ma) ages for the Buckskin Peak sample indicate rapid cooling between the closure temperatures of hornblende (~500°C) and biotite (~300°C). Because of the rapid cooling the igneous age is also interpreted to be ~149 Ma, although it could be somewhat older if the pluton cooled slowly from its solidus to 500°C.

Intrusions Cutting the Galice Formation

Sills in Galice Formation. Two sills (D24 and D26) were dated from the basal part of the Galice Formation 10 km northeast of Gasquet. This locality is where a depositional contact with the Josephine ophiolite is exposed [*Harper*, 1983] and where detailed radiolarian biostratigraphy has been done [*Pessagno and Blome*, 1990].

Sample D24 is a boudinaged sill within slaty argillites of the basal hemipelagic sequence. The age spectrum for relict igneous hornblende is unusual in that an age gradient is evident from 130 to 150 Ma (Figure 11a). The regional metamorphic grade at this locality is only prehnite-pumpellyite facies, so the age gradient cannot be the result of slow cooling through the closure temperature of unexolved igneous hornblende (~500°C). The 150 to 130 Ma gradient is probably the result of continued degassing of a low retention site in the hornblende during regional cooling. The igneous age for this sill is interpreted to be 150.5±1.4 Ma based on a plateau defined by the six highest temperature steps.

Sample D26 is a sill from the same locality as D24 but is located farther up section within thick bedded graywacke. The sill is largely recrystallized to metamorphic minerals and has fibrous extension veins related to formation of slaty cleavage in the Galice [*Harper*, 1992]. Relict igneous hornblende gives a well defined isochron age of 146.2±1.0 Ma (Figure 11b) which is considered to be an igneous age.

Lower Coon Mountain pluton. The Lower Coon Mountain pluton is a sill-like body that intrudes the lower part of the Galice Formation (Figure 2). Deep erosion has exposed both the floor and roof of the pluton, and the maximum thickness is >300 m [*Harper*, 1980]. The pluton consists almost entirely of clinopyroxene-rich ultramafic rocks [*Harper*, 1980; *Gray and Page*, 1985]. Minor gabbro, diorite, quartz diorite, and rare granodiorite occur along the margins and roof of the pluton. The pluton has contact metamorphosed the Galice Formation to biotite hornfels within a thin aureole [*Harper*, 1980].

A zircon age of ~142 Ma was reported by *Saleeby et al.* [1982] for a late granodiorite dike (J84z) intruding clinopyroxenite along the roof of the Lower Coon Mountain pluton. A second dike was sampled at this locality for U/Pb dating and consists of hornblende-biotite quartz diorite. Zircon data for this new sample (sample 2) are plotted on the concordia diagram of Figure 6a, along with samples 1 and 3 from the Grants Pass and Pony Peak plutons, respectively. Four size fractions from sample 2 yield lower and upper intercepts of 145 +1/-2 Ma and 755 +158/-174 Ma, respectively. The linear dispersion of data points away from the lower intercept, with increasing discordance correlating with increasing grain size, is indicative of contamination of ~145 Ma cognate zircon with significantly older zircon from either source domains (inheritance) or ascent/emplacement domains (entrainment). Thus the lower intercept of 145 +1/-2 for sample 2 is interpreted to be the igneous age of the sample.

An $^{40}Ar/^{39}Ar$ age spectrum for a K-feldspar (J84z) from the ~142 Ma granodiorite dike dated by *Saleeby et al.* [1982] is shown in Figure 12. It has early steps that increase in age until a plateau is reached. The plateau comprises 70% of the total ^{39}Ar released and gives an isochron age of 134.9±0.9 Ma. Closure temperatures are variable for K-feldspars but can be calculated from data obtained during the step heating [*McDougall and Harrison*, 1988]. In addition, it has recently been shown that a K-feldspar can have several discrete diffusion domains, each having its own closure temperature [*Lovera et al.*, 1989]. The calculated closure temperatures are ~300°C for the 135 Ma plateau and ~180°C for the initial steps. The plateau age is interpreted as the time that the Lower Coon Mountain pluton cooled below ~300°C.

In summary, the Lower Coon Mountain pluton was intruded at ~145 Ma, followed by slow cooling to ~300°C at 135 Ma. The zircon and K-feldspar ages, along with widespread minor prehnite-pumpellyite facies recrystallization in the pluton, suggest intrusion of the pluton during Nevadan regional metamorphism.

In addition to the Lower Coon Mountain pluton, the Pony Peak and Grants Pass plutons intrude the Galice Formation (Figure 2). These plutons also cut the Orleans thrust and are thus discussed in the following section.

AGE OF PLUTONS CUTTING THE ORLEANS THRUST

The Summit Valley, Pony Peak, and Grants Pass plutons cut the Orleans thrust (Figure 2). Only the Summit Valley has been studied in detail and is clearly syntectonic [*Griesau and Harper*, 1992; *Griesau*, 1992], but preliminary work discussed below for the Grants Pass and Pony Peak plutons suggests they too are syntectonic. Nevertheless, all three plutons contact metamorphose both the upper and lower plates or the Orleans thrust, suggesting intrusion occurred after thrusting but before the end of deformation.

Summit Valley Plutonic Complex

Contact relationships and lithology. The Summit Valley plutonic complex consists of numerous individual intrusions. The oldest phases are clinopyroxenite and hornblendite followed by hornblende gabbro and finally minor hornblende diorite, quartz diorite, and tonalite [*Norman*, 1984].

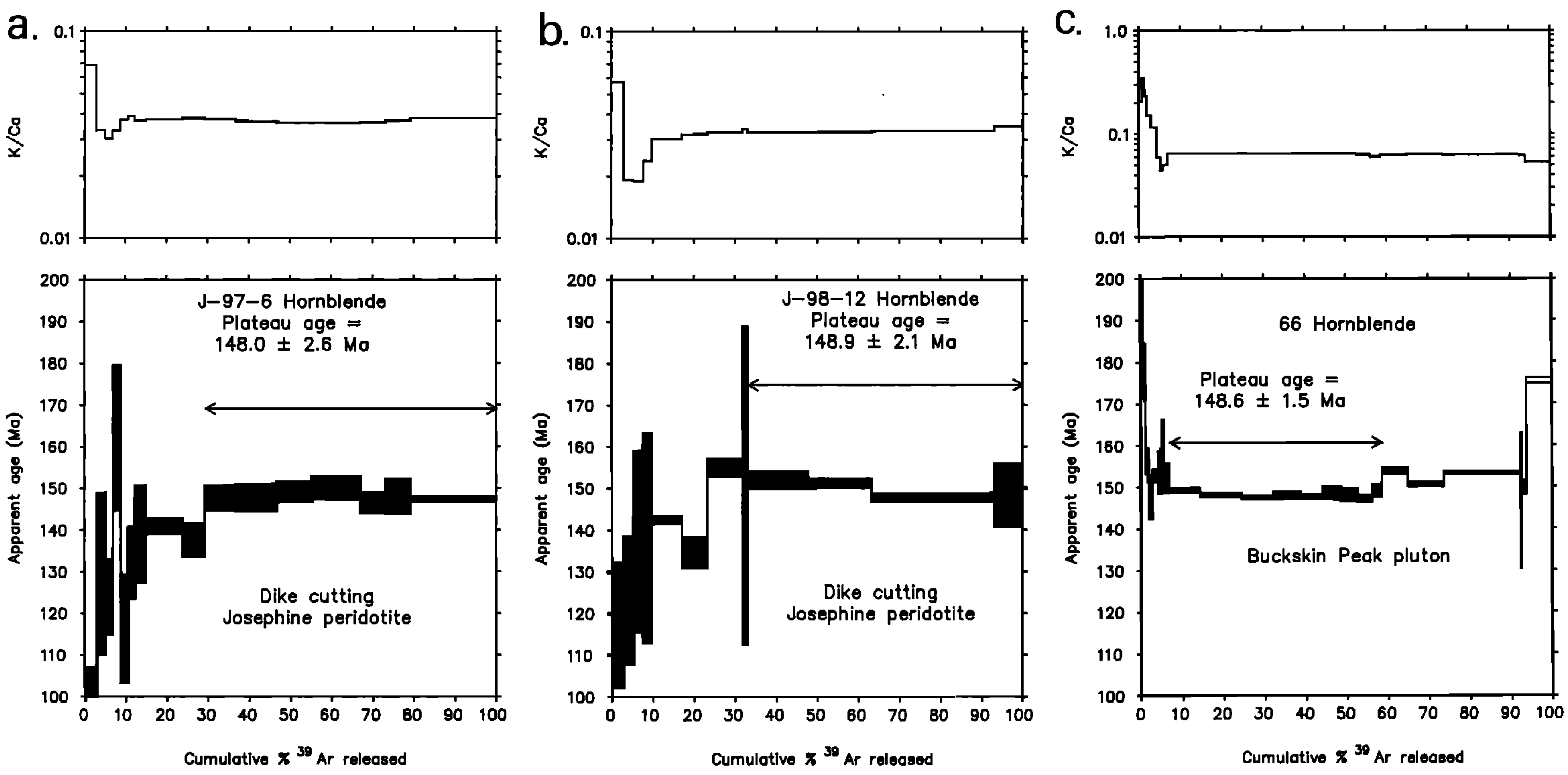


Fig. 10. (a) and (b) $^{40}Ar/^{39}Ar$ age spectra for dikes intruding the basal peridotite of the Josephine ophiolite, 12 and 14 km west of Cave Junction, Oregon. (c) $^{40}Ar/^{39}Ar$ age spectrum for hornblende from quartz diorite of the Buckskin Peak pluton.

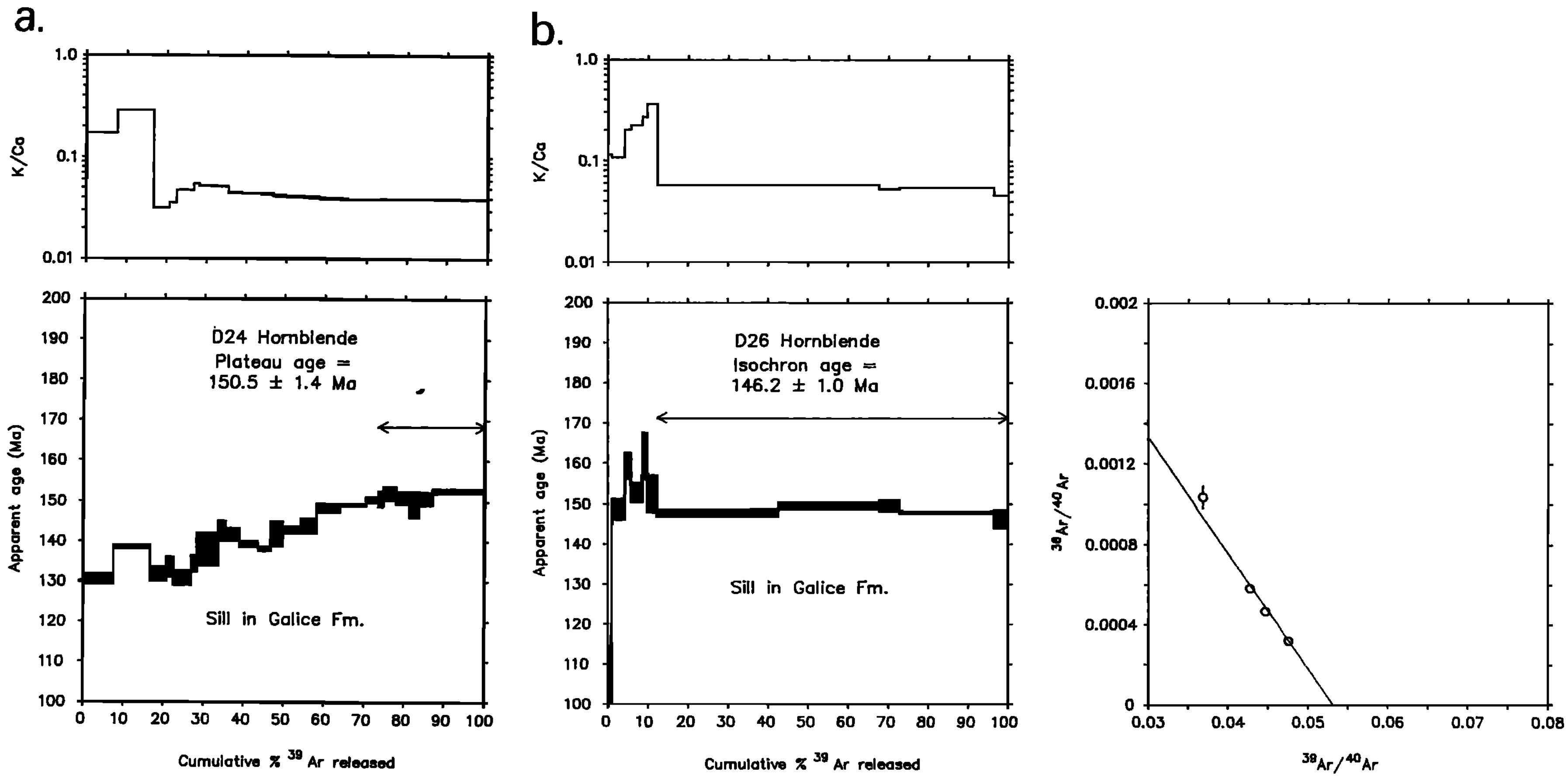


Fig. 11. (a) $^{40}Ar/^{39}Ar$ age spectrum for a metamorphosed and deformed sill in the basal part of the Galice Formation, 10 km northeast of Gasquet. See text for discussion. (b) $^{40}Ar/^{39}Ar$ age spectrum and isochron plot for a metamorphosed sill in the lower Galice Formation.

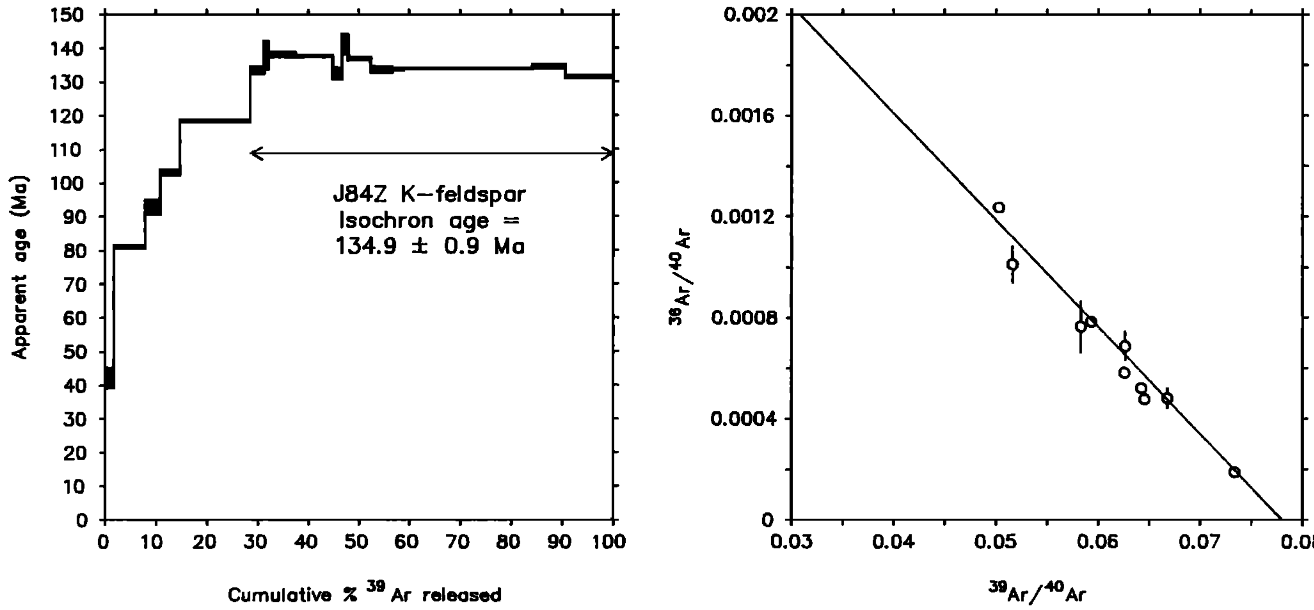


Fig. 12. $^{40}Ar/^{39}Ar$ age spectrum and isochron plot for a K-feldspar from a granodiorite dike in the Lower Coon Mountain pluton. The calculated closure temperature for this K-feldspar is ~300°C.

The Summit Valley plutonic complex intrudes and contact metamorphoses both the upper and lower plates of the Orleans thrust (Figure 2). The upper plate consists of serpentinized peridotite of the Rattlesnake Creek terrane, and the lower plate consists of the Lems Ridge olistostrome which underlies the Galice Formation in this area (Figure 4c). The contact aureole varies from static to strongly deformed [*Norman*, 1984; *Griesau*, 1992].

Minor high-temperature, crystal-plastic deformation is extensive in the plutonic complex, with hornblende and/or plagioclase partially recrystallized to very fine-grained aggregates. Highly strained rocks occur in shear zones ranging from <1 cm to >300 m in width, located both within and along the margins of the plutonic complex. The shear zones consist of foam textured, upper amphibolite-facies mylonites (plagioclase ± hornblende ± clinopyroxene) that are locally cut by centimeter-scale lower amphibolite-facies mylonites. Crosscutting relationships show that intrusion and crystal-plastic deformation overlapped in time [*Griesau*, 1992]. Late stage metamorphism and deformation includes faults having epidote or prehnite slickenfibers, widespread minor static recrystallization of plutonic rocks, and greenschist-facies assemblages in late stage microdiorite dikes [*Norman*, 1984]. These relationships indicate that retrogressive metamorphism and localized deformation occurred as the pluton cooled. The mylonitic shear zones and the late slickenfiber faults are highly variable in orientation but appear to have all formed under a stress field similar to that which produced structures in the Galice Formation [*Griesau and Harper*, 1992]. The largest shear zone (~300 m wide), which is far larger than any other, has a top-to-the-west sense of shear [*Griesau*, 1992].

Geochronology. Sample 4 is a hornblende gabbro collected from a large road cut where at least four different crosscutting intrusions of gabbro are evident. Zircon data for two fractions from this sample are plotted in Figure 6b and are externally concordant at 150±1 Ma, which is interpreted to be the igneous age.

A late stage, coarse-grained pegmatitic gabbro having elongate hornblende (SV-1h) was sampled from this same exposure for $^{40}Ar/^{39}Ar$ dating. In thin section, an igneous foliation is defined by the preferred orientation of hornblende and zoned plagioclase. Much of the plagioclase is recrystallized to fine-grained aggregates indicating high-temperature solid-state deformation, but the hornblende is undeformed. The hornblende yielded a plateau representing 95% of the released Ar, and steps from the plateau define an isochron age of 144.1±0.4 Ma (Figure 13a). The high-temperature deformation evident in this sample suggests it cooled relatively slowly, and thus the igneous age of this late stage dike is bracketed between 144 Ma (hornblende cooling age) and 150 Ma (zircon age on older gabbro from the same outcrop). A pluton as small as the Summit Valley would be expected to cool rapidly to the ambient temperature of the wall rocks, but continual addition of heat from successive intrusive phases of the plutonic complex apparently kept temperatures elevated above 500°C until 144 Ma.

In summary, the U/Pb zircon age indicates at least one phase of the Summit Valley plutonic complex was intruded at 150 Ma, and the plutonic complex remained above ~500°C until 144 Ma. Because the plutonic complex cuts the Orleans thrust and contact metamorphoses both the upper and lower plate, it is interpreted to postdate displacement on the thrust. Strong evidence for syntectonic intrusion, however, indicates that deformation continued after 150 Ma. The high-temperature shear zones in the plutonic complex are constrained by the age data to have been active between ~150 and 144 Ma, whereas faults with epidote and prehnite slickenfibers must have formed after 144 Ma [*Griesau and Harper*, 1992; *Griesau*, 1992]. The age data for the plutonic complex, along with its retrogressive metamorphic and deformation history, suggest it was intruded and cooled during Nevadan regional metamorphism.

Pony Peak Pluton

Contact relationships and lithology. The Pony Peak pluton (Figure 2) also intrudes and contact metamorphoses the upper

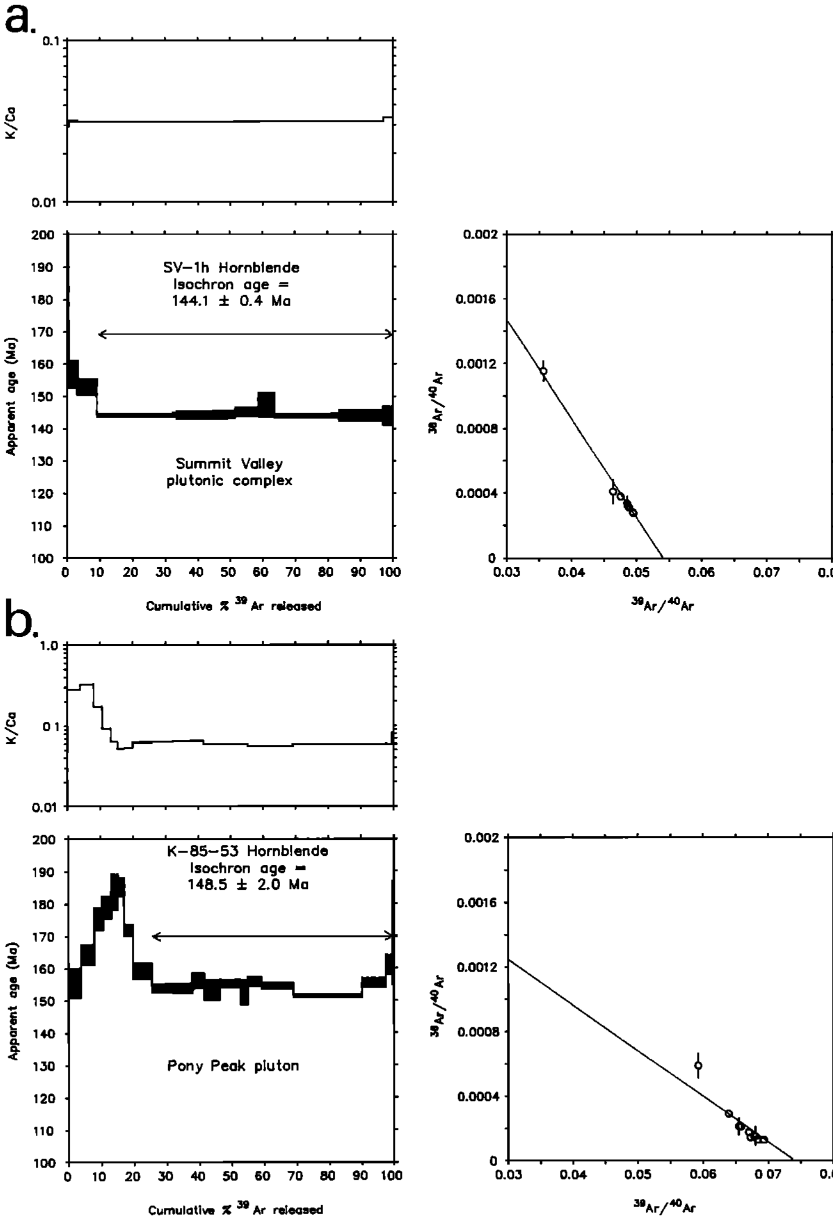


Fig. 13. (a) $^{40}Ar/^{39}Ar$ age spectrum and isochron plot for hornblende from a late stage pegmatitic gabbro in the Summit Valley plutonic complex. (b) $^{40}Ar/^{39}Ar$ age spectrum and isochron plot for hornblende from a tonalite of the Pony Peak pluton. (c) $^{40}Ar/^{39}Ar$ age spectrum and isochron plot for hornblende from a quartz diorite dike along the margin of the Pony Peak pluton. U/Pb data for zircon from this sample are shown in Figure 6a (sample 3).

and lower plates of the Orleans thrust [*Wagner and Saucedo*, 1987; S. *Cashman*, written communication, 1986]. The upper plate consists of ophiolitic melange of the Rattlesnake Creek terrane and slaty metasedimentary rocks of the Western Hayfork terrane(?). The lower plate consists of the Galice Formation which has been converted to a lineated biotite-andalusite schist in the contact aureole.

The Pony Peak pluton is predominantly tonalite, but ranges from gabbro to granodiorite (C.G. Barnes, personal communication, 1989). Syntectonic intrusion is evident in exposures on the north side of Pony Peak where numerous small mylonitic shear zones in hornblende diorite are truncated by undeformed dikes and by a biotite-hornblende tonalite phase of the pluton [C. Barnes and G. Harper, unpublished field data, 1991]. The shear zones consist of very fine-grained polygonized plagioclase + hornblende, similar to high-temperature mylonites of the Summit Valley plutonic complex. Syntectonic intrusion is also suggested by the nature of the contact aureole in that the Galice Formation has been converted to a strongly foliated and lineated schist rather than to a hornfels. Deformation during retrogressive metamorphism is suggested by muscovite rims around andalusite porphyroclasts, both of which show textural evidence for rotation during ductile shear. The pluton has also been affected by low-grade metamorphism: much of the plagioclase is replaced by epidote + muscovite + albite, and biotite and hornblende are partially replaced by chlorite ± epidote. The metamorphic and deformation history of the pluton and its aureole suggests that it was intruded during Nevadan regional metamorphism.

The two samples collected for geochronology include a

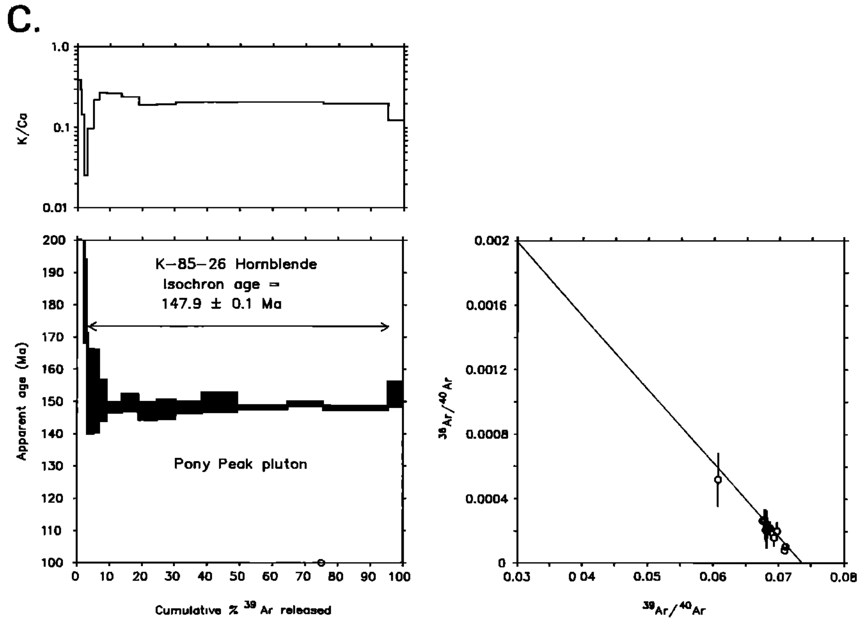


Fig. 13. (continued)

hornblende quartz diorite (K85-26) and a biotite-hornblende tonalite (K86-53). The latter sample is probably the same intrusive phase discussed above which cuts deformed diorite on Pony Peak, whereas the former is from a large dike along the southern margin of the pluton. Both samples are partially recrystallized to greenschist-facies assemblages. They also show evidence for minor crystal-plastic deformation of quartz, including strongly undulose extinction, subgrain structure, partial recrystallization, and sutured grain boundaries.

Geochronology. Zircon data for the Pony Peak pluton (sample 3, K85-26) are discordant and dispersed off concordia in a pattern indicative of a Proterozoic-dominated zircon population within a Late Jurassic magmatic system (Figure 6a and Table 2). On the concordia diagram of Figure 6a. the data from five fractions do not form a strong linear array, suggesting heterogeneity in the inheritance/entrainment component. The fraction of zircon that is fine, highly faceted, euhedral, and clean plots very close to concordia and thus strongly constrains the lower intercept. The nonlinearity of the array, however, leads to large uncertainties in the upper intercept. The lower intercept is 146±3 Ma and is considered an approximate igneous age. The poorly constrained upper intercept age of 1555 +598/-336 indicates a Proterozoic age contaminant.

The $^{40}Ar/^{39}Ar$ age spectra for hornblende from the zircon sample (K85-26) and a second sample (K85-53) both show anomalously old apparent ages for the initial steps, probably due to nonatmospheric trapped Ar. The remaining steps of both samples, comprising >80% of the total ^{39}Ar released, yield isochron ages of 147.9±0.1 and 148.5±2.0 Ma for samples K85-26 and K85-53 (Figure 13b and 13c), respectively. The 147.9 Ma age for K85-26 is interpreted to be an igneous age as well as a hornblende cooling age because the sample is from a dike along the margin of the pluton and thus probably cooled rapidly. This interpretation is consistent with the 146±3 Ma zircon age obtained for this sample.

In summary, the data suggest an igneous age of ~148 Ma for the Pony Peak pluton, at least for the younger phase. The pluton must postdate virtually all of the >40 km displacement on the Orleans thrust because the pluton cuts the thrust and contact metamorphoses both the upper and lower plates. Nevertheless, the pluton is apparently syntectonic as evident from the dynamic contact aureole and high-temperature deformation in the pluton, implying that Nevadan deformation continued after movement on the Orleans thrust ceased.

Grants Pass Pluton

Contact relationships and lithology. The Grants Pass pluton has intruded the Orleans thrust (Figure 2) and contact metamorphosed both the upper and lower plates [*Wells et al.*, 1940; *Snoke*, 1977; *Mack*, 1982]. In the lower plate, the Galice Formation has been converted to a fine-grained lineated muscovite-biotite-andalusite schist. Andalusite porphyroblasts have sigmoidal inclusion trails and asymmetric tails suggesting syntectonic growth.

The Grants Pass pluton is mostly quartz diorite and quartz monzonite, but varies from diorite to granodiorite [*Hotz*, 1971, *Barnes et al.*, 1992]. Much of the pluton has an igneous foliation and lineation, and subsolidus fabrics are locally present. Intrusion overlapping with deformation was noted at several localities, but it is particularly evident in exposures on the Applegate River, along the southwest margin of the pluton: quartz diorite having an igneous foliation and flattened xenoliths is cut by granitic dikes, some of which have mylonitic fabrics; these rocks are in turn cut by fine-grained mafic dikes that have been sheared parallel to their margins, resulting in lineated amphibolite having asymmetric porphyroclasts and shear bands. As with the Summit Valley and Pony Peak plutons, the preliminary observations on the Grants Pass pluton shows that it postdates major displacement on the Orleans thrust yet has a dynamic contact aureole and was internally deformed during the period of magma intrusion.

Geochronology. A uniform, inclusion-free domain of a granodiorite exposed on the Rogue River was sampled for zircon dating. The sample has a weak igneous foliation, and only minor solid-state deformation is evident (kinked biotite and microstructures in quartz). U/Pb age data from four zircon fractions are plotted on the concordia diagram (sample 1, Figure 6a). The data are dispersed away from the lower intercept of 139±2 Ma and project to an upper intercept of 1792 +168/-189. These data are clearly indicative of an inheritance/entrainment discordance mechanism, although the imperfection of the linear array suggests isotopic heterogeneity in the contaminant zircon. The proximity of the data for the fine fraction to concordia strongly constrains the lower intercept at 139±2 Ma, which is taken to be a close approximation to the igneous age of the sample. This age is concordant with a previous K/Ar hornblende age of 139±4 Ma from a quartz monzonite located several km to the northwest [*Hotz*, 1971].

Summary. The 139 Ma igneous age for the Grants Pass pluton makes it the youngest intrusion in the western Klamath Mountains. Nevertheless, it appears to be syntectonic and shows the same structural relationships as the Summit Valley and Pony Peak plutons. All three plutons cut the Orleans thrust and contact metamorphose both the upper and lower plates, show clear evidence for deformation overlapping in time with intrusion, and have dynamic contact aureoles in the lower plate.

The Grants Pass and Pony Peak plutons, along with the Lower Coon Mountain pluton, all show the effects of inheritance and/or entrainment of Proterozoic zircon (Figure 6a). The contaminant zircon may have been inherited during melting, or it may have been entrained from the Galice Formation which has a detrital zircon population dominated by Proterozoic species [*Miller and Saleeby*, 1987].

DISCUSSION

Figures 14 and 15 summarize the geochronologic data and illustrates the relatively short time interval between ophiolite generation and ophiolite emplacement. Intrusion of calc-alkaline magmas into the Josephine ophiolite and Galice Formation began at ~150 Ma, and the ages of these intrusions tightly constrain the timing of Galice deposition, displacement on the basal and roof thrusts, and regional metamorphism and deformation.

The Josephine Ophiolite

The revised 162±1 Ma U/Pb zircon age and 165±3 Ma $^{40}Ar/^{39}Ar$ hornblende age shows that the Josephine ophiolite in the study area is ~5 Ma older than previously determined [*Saleeby et al.*, 1982]. This revised age is similar to the 161±3 Ma and 165±3 Ma $^{40}Ar/^{39}Ar$ hornblende ages and 164±1 Ma zircon age [*Wright and Wyld*, 1986] for the Devils Elbow remnant of the ophiolite (Figure 4d). In the absence of additional ages on the Josephine ophiolite, the end of seafloor spreading is only constrained between 162 Ma and ~155 Ma (Figure 14). Extensional tectonics, however, appears to have continued until at least 157 Ma, based on the occurrence of Kuroko-type massive sulfide deposits in the uppermost Rogue Formation (Figure 4a) [*Koski and Derkey*, 1981; *Sillitoe*, 1982; *Sawkins*, 1990].

The Galice Formation

The age of the hemipelagics and flysch overlying the Josephine ophiolite is bracketed by the 162 Ma age of the ophiolite and by crosscutting calc-alkaline intrusions as old as 151 Ma (Figure 14). Arc volcanism was taking place during this interval as recorded by tuffaceous detritus in the hemipelagic sequence [*Pinto-Auso and Harper*, 1985], local volcaniclastic members in the Galice Formation (Figure 4), and abundant volcanic detritus in the Galice flysch [*Harper*, 1983; 1984].

The beginning of flysch sedimentation is marked by the first appearance of graywackes interbedded with radiolarian argillite ~50 m above the Josephine ophiolite (base of the Galice Formation, *sensu stricto* [*Pessagno and Blome*, 1990]). New radiolarian faunal data along with ammonite and bivalve fossils indicate these basal wackes are middle Oxfordian in age and correlative with the lowermost strata of the type Galice Formation in the Rogue Valley subterrane (Figure 4) [*Pessagno and Blome*, 1990]. The underlying hemipelagic strata are lower to middle Oxfordian, and rare strata within Josephine pillow lavas are upper Callovian [*Pessagno and Blome*, 1990]. These results imply that the earliest flysch sedimentation in both the Smith River and Rogue Valley subterranes began after 157 Ma (zircon age on Rogue Formation [*Saleeby*, 1984]).

The Galice flysch was largely derived from older rocks of the Klamath Mountains [*Snoke*, 1977; *Harper*, 1983, 1984; *Wyld*, 1985]. Previous workers have suggested that the source area was formed by uplift related to a Middle Jurassic (pre-Nevadan) orogeny [*Harper and Wright*, 1984; *Wright and Fahan*, 1988]. The new age data, however, suggest Galice flysch deposition was related to Nevadan thrusting: (1) the age of the flysch (~157 to 150 Ma) overlaps in time with movement on the basal and roof thrusts, (2) both the Smith River and Rogue Valley subterranes show a change from pelagic-hemipelagic sedimentation to flysch sedimentation during the middle Oxfordian, suggesting a sudden increase in terrigenous sediment, and (3) uplift of the source area for the flysch at 155-150 Ma is indicated by numerous hornblende and mica cooling ages in the central Klamath Mountains [*Donato* and Lanphere, 1992; *Saleeby* and *Harper*, 1993; *Hacker et al.*, 1993]. These data suggest the Galice flysch was a synorogenic deposit shed off the advancing Orleans thrust sheet. In this context, the Galice flysch could be viewed as a trench-like deposit related to underthrusting of the Josephine ophiolite.

Age and Displacement Rates of the Roof and Basal Thrusts

One of the main results from this study is that both the Madstone Cabin (basal) and Orleans (roof) thrusts underwent most of their displacement between ~155 and 150 Ma (Figure 14). Between 150 and 139 Ma the thrusts, as well as the Western Klamath terrane as a whole, were "stitched" together by calc-alkaline intrusives (Figure 14).

The field and geochronologic data for the amphibolite sole, along with new156 to 161 Ma zircon ages for deformed plutonic rocks of the Chetco complex [*Yule and Saleeby*, 1993], show that most of the penetrative deformation associated with the basal thrust took place under amphibolite-facies conditions between ~156 and 151 Ma. The thin and discontinuous nature of phyllonites suggests continued minor displacement on the basal thrust occurred during retrogressive greenschist-facies (150 to at least 146 Ma). The transport direction for the Madstone Cabin thrust is north-northeast for both amphibolites and phyllonites, with a top-to-the-northeast sense of shear [*Cannat and Boudier*, 1985; *Harper et al.*, 1990; *Grady*, 1990]. Large post-Nevadan clockwise rotations have affected the northern Klamath Mountains [Schultz and Levi, 1983; *Bogen*, 1986], implying that displacement was top-to-the-northwest at the time of thrusting. A minimum offset of ~12 km across the Madstone Cabin thrust is evident from its surface outcrop, yielding a minimum displacement rate of 2.4 mm/year for 155-150 Ma. As discussed above, it is likely that the Pearsoll Peak thrust (Figure 2) is a continuation of the Madstone thrust, which would yield a combined minimum offset of 42 km and a minimum thrusting rate of 8.4 mm/year.

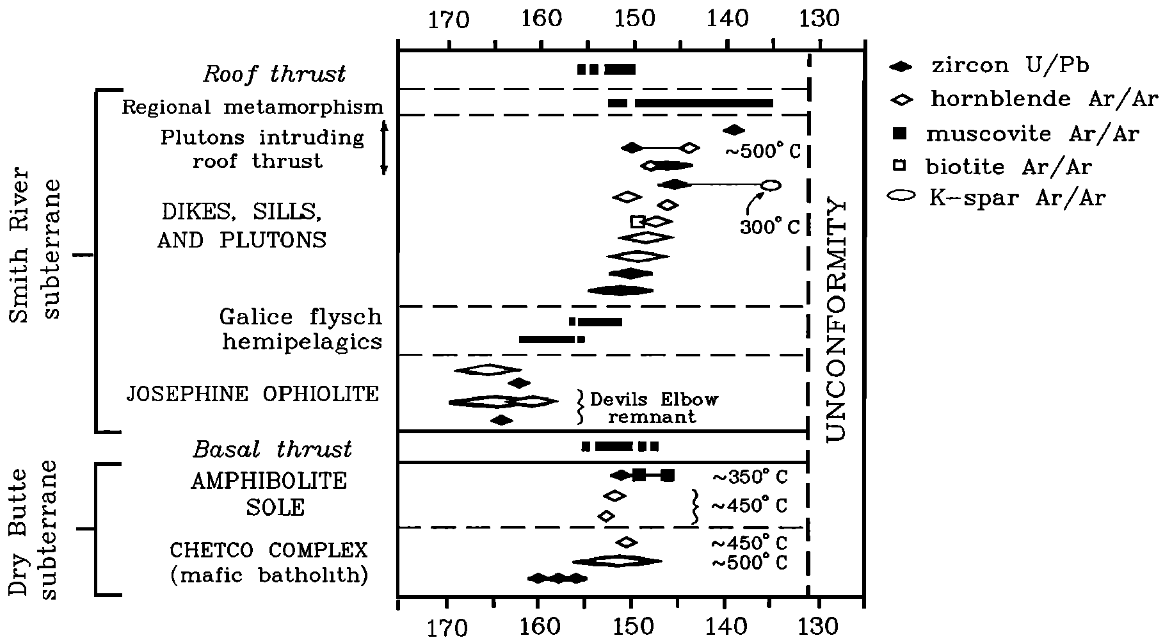


Fig. 14. Summary of $^{40}Ar/^{39}Ar$ and zircon ages in the Western Klamath terrane, interpreted as igneous ages except where closure temperatures are shown. The dikes, sills, and plutons section includes two zircon ages (151±3 and 150±2 Ma) from *Saleeby et al.* [1982]. U/Pb zircon age for Devils Elbow remnant is from *Wright and Wyld* [1986] and zircon ages for the Chetco complex are from J. D. Yule (personal communication, 1992) and *Yule and Saleeby* [1993].

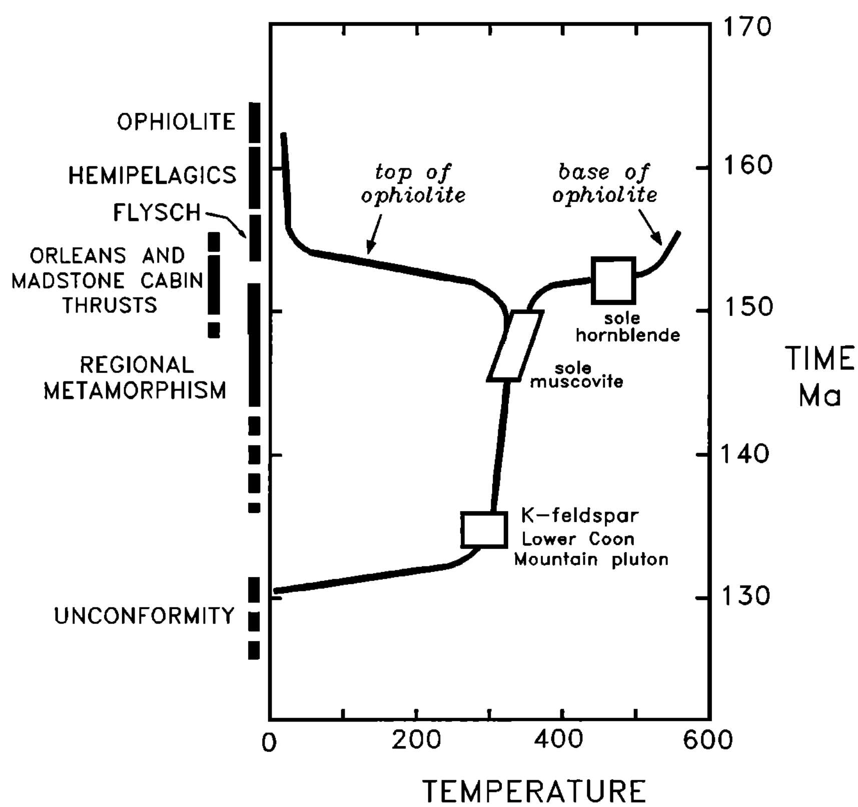


Fig. 15. Temperature-time history for the base and top of the Josephine ophiolite. Peak metamorphic temperatures are from *Harper et al.* [1988] and *Milby* [1990].

The minimum westward displacement along the Orleans thrust is >40 km based on its outcrop pattern (Figures 1 and 2), and Jachens et al [1986] have estimated >110 km displacement based on geophysical data. Structural evidence for west or northwest directed thrusting of the older Klamath terranes over the Western Klamath terrane has been presented by many workers [*Snoke*, 1977; *Harper*, 1980, 1992; *Gray*, 1985; *Wyld*, 1985; *Jones*, 1988; *Harper*, 1992]. The maximum age of the Orleans thrust is constrained by the truncation of the 161±2 Ma Wooley Creek batholith in the upper plate (Figure 1) [Jachens et al., 1986; *Barnes et al.*, 1986]. Virtually all movement on the Orleans thrust must have transpired by 150 Ma, because the 150 Ma Summit Valley plutonic complex, the 148 Pony Peak, and 139 Ma Grants Pass plutons cut the thrust and contact metamorphose both the upper and lower plates. Minor movement on the Orleans thrust may have continued until as late as 139 Ma because all three plutons show evidence for syntectonic, though not necessarily synthrusting, intrusion. For 40 km of minimum offset, the minimum rate of westward displacement is 3.6 mm/year, but assuming thrusting did not begin before ~155 Ma yields a minimum rate of 8 mm/year. The minimum rate of displacement would be 22 mm/year assuming >110 km of offset as inferred by Jachens et al. [1986]. This latter rate falls well within the range of modern plate velocities.

Calc-Alkaline Intrusions and Deformation

The new and previous ages [*Dick*, 1976; *Saleeby et al.*, 1982; *Wright and Fahan*, 1988] for calc-alkaline dikes, sills, and plutons in the Smith River subterrane define a narrow time interval of 151 to 145 Ma (Figures 2 and 15), except for the 139 Ma Grants Pass pluton. In addition, the Bear Mountains complex in the upper plate of the Orleans thrust (Figure 2) was intruded in two distinct pulses dated at 149±2 and 153±2 Ma [*Saleeby et al.*, 1982]. *Harper and Wright* [1984] divided the calc-alkaline intrusions into pre-Nevadan (153 to 150 Ma) and post-Nevadan (147 to 140 Ma). The new geochronologic data and field observations suggest that all the intrusions are syn-Nevadan.

The numerous dikes and sills intruding the Josephine ophiolite and Galice Formation were previously considered to be pre-Nevadan because they are regionally metamorphosed and many are deformed [*Harper*, 1980; *Saleeby et al.*, 1982; *Harper and Wright*, 1984; *Wyld*, 1985]. Deformation must have started by ~150 Ma, however, because some of the dikes are intruded into small precleavage thrust faults in the Galice Formation [*Harper*, 1992]. This observation is consistent with evidence for syntectonic intrusion of the 150 Ma Summit Valley and 148 Ma Pony Peak plutons. In addition, because these plutons cut the Orleans thrust, the Nevadan orogeny must have been well underway by ~150 Ma.

The end of Nevadan deformation was previously estimated to be ~145 Ma because the ~147 Ma Glen Creek and Ammon Ridge plutons have static contact aureoles that overprint slaty cleavage in the Galice Formation [*Harper and Wright*, 1984; *Wright and Fahan*, 1988]. Similarly, the 145 Ma Lower Coon Mountain pluton has a static aureole. The following evidence, however, suggests that deformation continued at least locally after ~145 Ma: (1) the 146±1_Ma sill (D26) in the basal Galice Formation is regionally metamorphosed and contains extension veins related to penetrative deformation in the Galice Formation [*Harper*, 1992]; (2) faulting associated with epidote and prehnite slickenfibers continued in the Summit Valley plutonic complex after cooling below 500°C at 144 Ma [*Griesau and Harper*, 1992]; and (3) deformation and intrusion overlapped in the 139 Ma Grants Pass pluton. K-feldspar from the Lower Coon Mountain pluton indicates temperatures remained above ~300°C until 135 Ma (Figure 15).

The relationship between foliation in the Grants Pass contact aureole and slaty cleavage in the Galice Formation needs further study, but the White Rock pluton farther north, which has a similar structural setting and age (141 Ma K/Ar biotite), has a dynamic contact aureole that overprints slaty cleavage [Kays, 1992]. Pluton intrusion was associated with (1) folding of the slaty cleavage in the Galice (F_2 folds), locally associated with a crenulation cleavage and lineation, and (2) regional greenschist- to amphibolite-facies metamorphism [Kays, 1992]. F_2 folds are widespread in the Western Klamath terrane and have been interpreted by most workers as post-Nevadan [*Snoke*, 1977; *Harper*, 1980; *Norman*, 1984; *Wyld and Wright*, 1988; *Cashman*, 1988]. The structural relationships in the Grants Pass and White rock plutons, however, as well as the local presence of a crenulation cleavage sometimes associated with neocrystallization of mica [*Norman*, 1984; *Gray*, 1985; *Cashman*, 1988], suggests that the F_2 folds are the result of late stage Nevadan deformation. The amount of structural work on the 151 to 139 dikes and plutons varies greatly, and additional work is needed to fully characterize these intrusions and their contact aureoles as has been done in the Sierra Nevada foothills [*Paterson and Tobisch*, 1988; *Paterson et al.*, 1991]. Nevertheless, the plutons showing evidence for syntectonic intrusion (Summit Valley, Pony Peak, Grants Pass, White Rock) are all intruded into the Orleans thrust, whereas those reported to have static contact aureoles (Lower Coon Mountain, Ammon Ridge, Glen Creek) are located far from the thrust. This pattern suggests that strain may have been partitioned into the upper part of the Josephine-Galice thrust sheet, perhaps related to minor post-150 Ma movement on the Orleans thrust.

Thermal History

The age constraints from geochronology (Figure 14) and crosscutting relationships discussed above were used along with $^{40}Ar/^{39}Ar$ ages to construct a thermal history for the top and base of the Josephine ophiolite (Figure 15). Critical $^{40}Ar/^{39}Ar$ cooling ages for the metamorphic sole and for K-feldspar from the Lower Coon Mountain are shown. The curve for the top of the ophiolite is for the vicinity of Gasquet, where the peak metamorphism of ~330°C is indicated by metamorphic mineral assemblages [*Harper et al.*, 1988] and fluid inclusion data [*Milby*, 1990]. The K-feldspar age shows that the area remained above ~300°C until 135 Ma (Figure 15), at least in the vicinity of the Lower Coon Mountain pluton.

The most striking aspect of the cooling curves is that during the major displacement on the Orleans and Madstone Cabin thrusts (~155 to ~150 Ma) there appears to have been a very steep thermal gradient between the amphibolite sole and the top of the ophiolite. After ~150 Ma, temperatures appear to have homogenized throughout the Josephine-Galice thrust sheet (Figure 15).

The source of the heat to produce the amphibolite sole may have come from the upper mantle rocks of the ophiolite, but there is no inverted metamorphic gradient beneath the basal thrust [*Harper et al.*, 1990]. It appears more likely that the heat came from the footwall as the ophiolite was thrust over the active Rogue-Chetco arc [*Dick*, 1976].

The Nevadan Orogeny

The previously estimated timing of the Nevadan Orogeny in the western Klamath Mountains was 150 to 145 Ma [*Saleeby et al.*, 1982; *Harper and Wright*, 1984; *Wright and Fahan*, 1988]. This age range brackets the timing of penetrative deformation (slaty cleavage and associated folding) characteristic of the Galice flysch as well as correlative formations in the Sierra Nevada foothills. The new geochronologic data are consistent with the 150-145 Ma age bracket for this deformation, but they also indicate a more protracted history of deformation that extended before and after slaty cleavage formation.

The Nevadan orogeny must have begun before 150 Ma since displacement on the roof and basal thrusts were essentially completed by 150 Ma. The maximum age for initiation of thrusting is bracketed by the 162 Ma Josephine ophiolite,156-161 Ma zircon ages of deformed intrusive rocks beneath the basal thrust [*Yule and Saleeby*, 1993]), and the 157 Ma zircon age of the Rogue Formation [*Saleeby*, 1984] structurally beneath the ophiolite (Figure 4a). Because these thrusts have such large displacements, it is clear that extensive crustal shortening occurred between ~155 and 150 Ma, apparently during deposition of the Galice flysch.

The end of the Nevadan orogeny is bracketed by relatively young plutons and an Early Cretaceous angular unconformity. Syntectonic intrusion of the 139 Ma Grants Pass pluton indicates that elevated temperatures and deformation persisted at least locally until ~140 Ma. Relationships evident around the ~140 Ma White Rock Pluton [Kays, 1992] show that folding of slaty cleavage was ongoing at ~140 Ma and suggest folding of slaty cleavage in the Galice flysch, evident throughout the Klamath Mountains [*Snoke*, 1977; *Harper*, 1980; *Gray*, 1985; *Wyld*, 1985], may have occurred at this time. We propose that such deformation, which occurred after formation of slaty cleavage in the Galice flysch and before cooling and uplift (i.e., between ~145 and 135 Ma), be referred to as "late phase Nevadan Orogeny." The timing of uplift is recorded by the 135 Ma cooling age for K-feldspar J84z (Figure 12). The end of the Nevadan orogeny is bracketed by an angular unconformity where Lower Cretaceous strata overlie the Galice Formation (Figures 14 and 15). These strata are as old as latest Hauterivian to Barremian near Cave Junction (Figure 1) [Imlay et al., 1959; *Nilsen*, 1984] and as old as late Valanginian in western outliers of the Galice Formation (Figure 1) [*Blake et al.*, 1985]. According to the Decade of North American Geology [*Palmer*, 1983] and *Harland et al.* [1989] time scales, late Hauterivian corresponds to ~125 Ma and ~133 Ma, respectively, whereas late Valanginian corresponds to ~132 and ~139 Ma.

We interpret these relationships to indicate that regional metamorphism and at least local deformation continued until ~135 Ma, or 5 to 10 Ma longer than what is generally accepted as the end of the Nevadan orogeny. The late deformation appears to be temporally continuous with the main phase of deformation and metamorphism, although it appears to be less intense and more localized. Similarly, structures and textures typical of peak Nevadan conditions are documented to have extended to as late as ~135 Ma in the southern foothills of the Sierra Nevada [*Saleeby et al.*, 1989a; *Tobisch et al.*, 1989]. In addition, ductile deformation features are well developed within and around 143 to 138 Ma plutons that cut main phase Nevadan structures in the northern Sierran Foothills [*Saleeby et al.*, 1989b]. These relations suggest that regional deformation and metamorphism related to the Nevadan orogeny lasted until ~135 Ma along the entire ~1000-km-long Jurassic belt extending from the Sierran foothills through the Klamath Mountains. The late strains were inhomogeneous as shown by a number of 140 to 150 Ma plutons which cut main phase Nevadan structures or have static aureoles which overprint them [*Wright and Fahan*, 1988; *Saleeby et al.*, 1989b].

The age of unconformably overlying Lower Cretaceous sedimentary rocks is less than 10 Ma and possibly less than 5 Ma after uplift and cooling of the Josephine-Galice thrust sheet below ~300°C (135 Ma K-feldspar age). Such a rapid change from regional metamorphism to deposition of marine strata indicates rapid exhumation. This high rate is not compatible with exhumation simply by uplift and erosion followed by subsidence and deposition. *Extension* is considered a more likely mechanism since it can produce both rapid exhumation and lowering of the surface height [*England Molnar*, 1990]. If this is correct, the end of the Nevadan orogeny (as defined in this paper) represents a major change from compressional to extensional tectonics.

CONCLUSIONS

The primary geochronologic results are summarized in Figure 15. The major conclusions are the following:

1. The age of the Josephine ophiolite in the study area is ~162 Ma, based on a 162±1 Ma U/Pb zircon age and a 165±3 Ma $^{40}Ar/^{39}Ar$ hornblende age (Figure 15). The 161±3 and 165±5 Ma $^{40}Ar/^{39}Ar$ hornblende ages for the Devils Elbow remnant of the ophiolite are consistent with a previous U/Pb zircon age of 164±1 Ma [*Wright and Wyld*, 1986].

2. The Madstone Cabin (basal) and Orleans (roof) thrusts of the Josephine ophiolite were both active between ~155 Ma and 150 Ma. Displacement on both thrusts essentially ended by 150 Ma.

3. In the amphibolite sole, amphibolite facies metamorphism ended by ~151 Ma, syntectonic pegmatites were intruded at 151±1 Ma, and retrogressive greenschist facies metamorphism and local deformation took place between 151 Ma and ~146 Ma or later.

4. The minimum displacement rate on the basal (Madstone Cabin) thrust is 2.4 mm/year, and is 8.4 mm/year assuming the Pearsoll Peak thrust is a northern continuation of the Madstone Cabin thrust. For the roof (Orleans) thrust, calculated minimum displacement rates are >8 mm/year for a minimum of 40 km offset, and >22 mm/year assuming the >110 km offset inferred by Jachens et al. [1986] is correct.

5. From ~155 to 151 Ma there was a strong thermal gradient between the basal thrust (amphibolite facies) and the top of the ophiolite. These gradients were gone by ~150 Ma, after which time low-grade regional metamorphism, penetrative deformation, and intrusion of calc-alkaline magmas affected the entire Western Klamath terrane.

6. Calc-alkaline magmas were intruded into the amphibolite sole, Josephine ophiolite and overlying Galice Formation, and the roof thrust between 151 to 139 Ma (Figure 15), consistent with results from previous studies [*Dick*, 1976; *Saleeby et al.*, 1982; *Wright and Fahan*, 1988]. Three plutons, ranging in age from 150 to 139 Ma, show evidence for syntectonic intrusion.

7. The Nevadan orogeny in the western Klamath Mountains was of longer duration (~155 to 135 Ma) that previously estimated (~150 to 145 Ma). The early phase of the Nevadan orogeny (155 to 150 Ma) involved major crustal shortening by thrust faulting and deposition of the Galice flysch. The late phase of the Nevadan orogeny (~145-135 Ma) involved

continued regional metamorphism and at least local deformation (Figure 15).

8. The Josephine ophiolite and other rocks of the Western Klamath terrane were exhumed and covered by unconformably overlying strata within only 5 to 10 Ma. Such rapid rates suggest unroofing by extensional tectonics.

The tectonic framework incorporating these results is shown in Figure 16. A modified version of the back-arc basin model of *Harper and Wright* [1984] is shown for generation of the Josephine ophiolite. As discussed by *Wyld and Wright* [1988], the newer older ages for the Josephine ophiolite are inconsistent with some aspects of this model: (1) rifting in the western Klamath Mountains started ~5 Ma prior to cessation of magmatism in the "remnant arc" exposed east and structurally above the Josephine ophiolite (Figure 16), and (2) the ophiolite now appears to be 2 to 5 m.y. older than the known age range of the structurally underlying "coeval" arc complex (Rogue-Chetco arc). The age overlap between the "remnant arc" and ophiolite

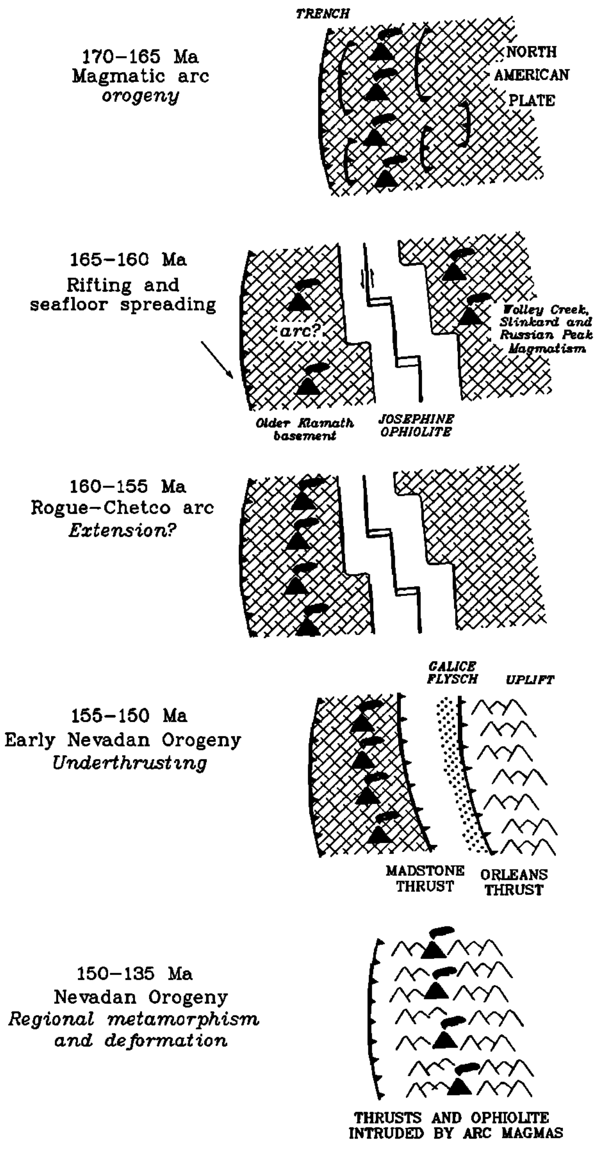


Fig. 16. Tectonic model for the generation and emplacement of the Josephine ophiolite, modified from *Harper and Wright* [1984], *Harper et al.* [1985], and *Wyld and Wright* [1988].

generation may have been the result of spreading along an intra-arc transform as shown in Figure 16 [*Harper et al.*, 1985; *Wyld and Wright*, 1988; *Saleeby*, 1990]. Alternatively, rifting may have initiated in the fore-arc region. In any event, the presence of older Klamath basement and rift edge facies in the Western Klamath terrane (Figures 4a, 4c, and 4d) implies that the Josephine basin formed by rifting of a magmatic arc built on the western edge of the North American plate [*Wyld and Wright*, 1988; *Yule et al.*, 1992].

The early phase of the Nevadan orogeny (~155-150) Ma involved crustal shortening by very large displacement on both the roof and basal thrusts (Figure 16). The large offset and very high displacement rate on the roof (Orleans) thrust, coeval displacement along the basal (Madstone Cabin) thrust, and contrasting thrust directions [*Harper*, 1992] suggest the ophiolite may have acted as a microplate between western North America and the Rogue-Chetco island arc from ~155-150 Ma (Figure 16).

We infer that underthrusting along the roof (Orleans) thrust resulted in crustal thickening and uplift of older terranes to the east (Figure 16). This uplift is recorded by (1) numerous ~155-150 Ma Ar/Ar hornblende cooling ages in the central Klamath Mountains [*Donato* and Lanphere, 1992; *Saleeby* and *Harper*, 1993; *Hacker et al.*, 1993], and (2) the onset of flysch sedimentation at ~157 Ma in both the Smith River and Rogue River subterranes (Figure 4). Flysch deposition is inferred to have taken place in a trench-like basin in front of the Nevadan highland (Figure 16), resulting in the flysch being progressively overridden by the advancing thrust sheet.

Strong thermal gradients were created during ~155-150 thrusting as a consequence of (1) underthrusting of an active arc complex (Figure 4a) beneath the relatively cold Josephine ophiolite to produce the amphibolite sole, and (2) underthrusting of the Josephine ophiolite beneath the western edge of North America, perhaps forming the inverted metamorphic gradient beneath the inferred eastern continuation of the roof thrust (Condrey thrust) [Jachens et al., 1986; *Saleeby* and *Harper*, 1993]. By 150 Ma, thrusting on both the roof and basal thrusts was essentially complete. Thermal gradients dissipated and regional low-grade metamorphism commenced. The Josephine ophiolite and underlying Rogue-Chetco arc were now accreted to the North American plate and buried to depths of 5 to 10 km. By 150 Ma, crustal shortening by thrusting left the ophiolite situated over the axis of arc magmatism, resulting in intrusion of numerous 150 to 139 Ma dikes and plutons (Figure 16). Penetrative deformation of the Galice flysch (e.g., slaty cleavage) took place from ~150 to 145 Ma, and regional metamorphism and at least local deformation continued until 135 Ma, at which time there was very rapid exhumation followed by deposition of unconformably overlying Lower Cretaceous strata (~130 Ma; Figures 15 and 16).

The geochronologic data suggest major changes in tectonic style occurred at ~165 Ma (opening of Josephine basin), ~155 Ma (beginning of crustal shortening), and 135 Ma (rapid unroofing, probably by extension). In addition, subsequent reheating to ~250°C at ~112 Ma is evident from K-feldspar incremental heating data (G. D. Harper, M. Heizler, and M. Roden, unpulished manuscript, 1993). Correlation of these times with plate reconstructions and the Late Jurassic apparent polar wander path for North America, which is currently controversial [*Van der Voo*, 1992; *Hagstrum*, 1993], can potentially show whether periods of extension and orogeny correlate which changes in motion of the North American and Pacific plates.

Acknowledgments. Reviewers R.A. Schweickert, R. Drake, D. Cowan, and especially J.E. Wright provided many valuable comments which led to significant improvements. Calvin Barnes and E. Darby Dyar also provided helpful comments and discussions. We are grateful to Henry Dick for providing mineral separates for samples he had dated by the K/Ar method. Dave Harding and Kristen Grady provided samples from the amphibolite sole, and Sue Cashman provided samples from the Pony Peak pluton. Mark Harrison provided valuable discussions and generous access to $^{40}Ar/^{39}Ar$ geochronologic facilities at SUNY Albany. This research was funded by National Science Foundation grants EAR8518974 and EAR8722425 (Harper) and AE08415114 and EAR8318212 (Saleeby).

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(Received June 5, 1991;
revised July 1, 1993;
accepted July 20, 1993.)